Volume 29 SEPTEMBER, 1959 Number 3

Journal of Sedimentary Petrology

A Publication of the Society of Economic Paleontologists and Mineralogists
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FIGS. 1–7, SEPTEMBER, 1959

AXIAL AND MARGINAL SEDIMENTATION IN GEOSYNCLINAL BASINS[1]

JOHN L. KNILL
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ABSTRACT

During the formation of geosynclines, greywackes tend to be deposited by axially directed turbidity currents; at the same time, a marginal facies may be developed. The Lower Palaezoic Geosyncline of Wales (Great Britain) contains some stratigraphical successions which appear to have been derived by both axial and lateral supply; the Harlech Grits (Cambrian) and the Upper Llandovery greywackes (Silurian) were deposited in NE-SW trending basins which were margined by areas of shelf-sea. Similarly, the Crinan Grits (Dalradian) of Scotland are partly made up of sedimentary material derived from the sides of the geosynclinal basin. The extent of interdigitation of the marginally and axially derived material appears to be associated with the form of the geosyncline.

INTRODUCTION

The recent investigations into the origin and orientation of sedimentary structures formed by current flow have become of considerable palaeogeographical importance. Many of these studies have been carried out in greywacke or flysch-type deposits where palaeocurrent directions run along the major tectonic trend (Kuenen, 1957, 1958; Haaf, 1957). It is clear that much sedimentary material of this type was deposited by turbidity currents flowing along geosynclinal basins formed parallel to the later mountain chains. However, in many cases comparatively little palaeocurrent data are available for the marginal facies of the same sedimentary basins. Where such information is present, it can generally be demonstrated that the marginal facies is derived from the sides of the basins. This paper is concerned with the description of some examples of palaeogeographical patterns of this type from the British Isles.

THE LOWER PALAEOZOIC GEOSYNCLINE OF WALES

The development of the Welsh Lower Palaeozoic Geosyncline has been described in two classic papers by Jones (1938, 1955); the reader is referred to these papers for a complete discussion of the stratigraphy and structural evolution of this basin.

Lower and Middle Cambrian

The earliest deposits of the Lower Palaeozoic Geosyncline are a succession of Cambrian rocks, largely comprising grits, flags and slates, which crop out in several disconnected areas (fig. 1). The main outcrop is in the Harlech Dome, where most of the sequence consists of grits showing graded bedding and many typical greywacke structures; there are also intercalated flaggy groups. Palaeocurrent observations by Kuenen (1953) and Kopstein (1954) have suggested that the main direction of derivation of these sediments was from the south-southwest. However, Matley and Wilson (1946) and Woodland (1939) have demonstrated on stratigraphical and lithological grounds that the succession appears to have originated from the northeast. Some observations by the author suggest that both these opposing views are, in part, correct. Palaeocurrent directions in one group, the Barmouth Grits, suggest a southwesterly derivation while another group, the Rhinog Grits, appear to have been derived largely from the northeast. Isopachytes for the lower part of the Harlech Dome succession (base of the Llanbedr slates to the base of the Middle Cambrian Clogau Shales) indicate a marked thickening of the grits towards the northeast (fig. 2a). All the palaeocurrent directions observed in a small part (6 square miles) of the Harlech Dome are shown in figure 2b; these observations covered the complete grit succession up to the base of the Middle Cambrian Clogau Shales. This diagram demonstrates that most of the currents flowed with a general north-south trend. Some of the finer-

[1] Manuscript received February 14, 1959.

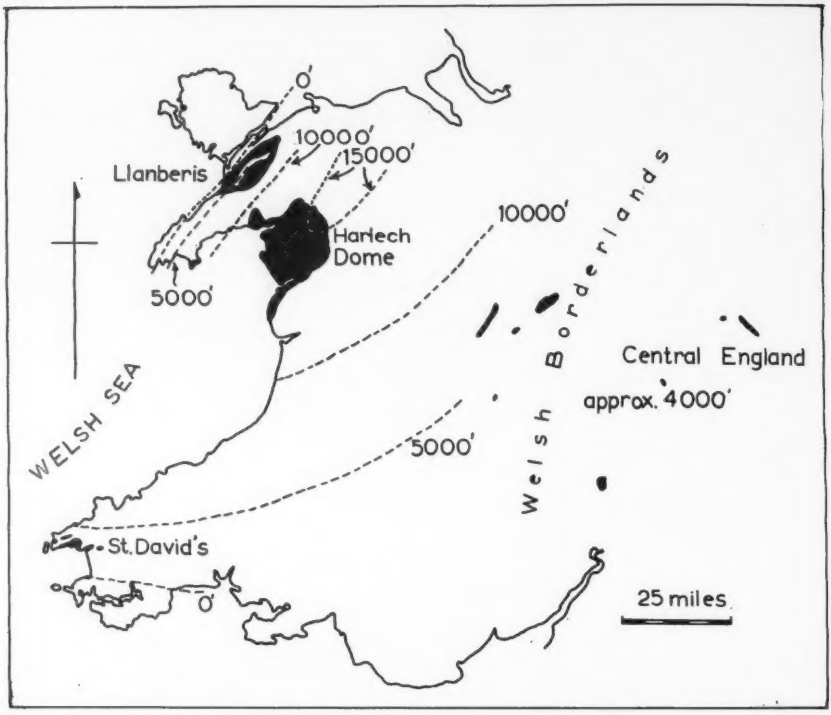

FIG. 1.—Distribution of the Cambrian of Wales and central England (outcrops shown in black). Isopachytes (in feet) for the Cambrian are based on Jones, 1955, Plate XVII.

grained, flaggy zones show more variation in direction of derivation and appear to have originated, in part, from the southeast. These results may appear, on first sight, to be rather anomalous; however, the picture becomes clearer when the Harlech outcrop is related to two other Cambrian areas at Llanberis and St. David's (fig. 1). Jones (1955, Plate XVII) has demonstrated that the general trend of the Cambrian isopachytes is NE-SW, with the thickest succession occurring in the Harlech area (fig. 1); the sequence thins noticeably towards the north-west and south. The Llanberis outcrop is largely composed of slates, with some grit intercalations, and "there are several gaps in the succession which indicate periodic uplift and possibly erosion during the Cambrian" (Jones, 1955, p. 325). Similarly, the St. David's sequence is

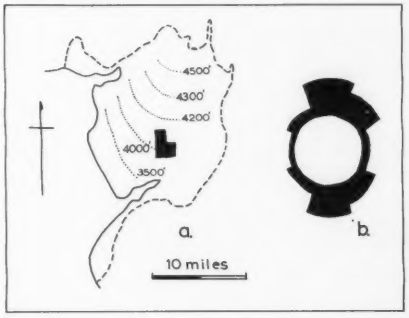

FIG. 2.—a. Isopachytes (in feet) for part of the Lower and Middle Cambrian succession of the Harlech Dome. b. Rose diagram of 80 palaeocurrent readings from the area shown in black in a. (The readings are measured outwards from the circle.)

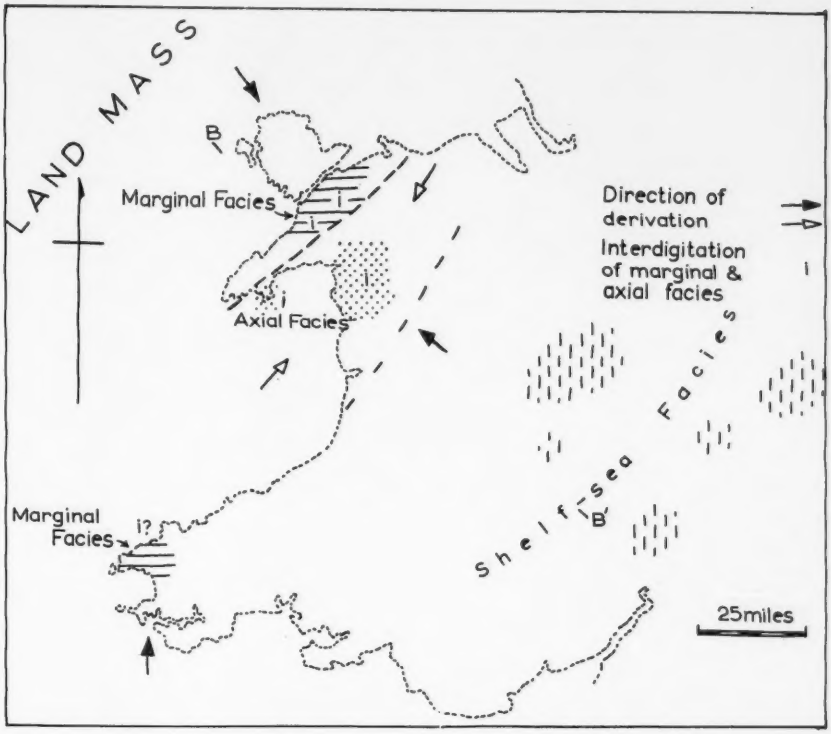

FIG. 3.—Palaeogeography of the Lower and Middle Cambrian. B-B′ is the line of section shown in figure 6.

both thinner and finer-grained than that at Harlech. Small Cambrian outcrops occur farther east in the Welsh Borderlands and Central England (fig. 1); these rocks show evidence of shallow-water conditions and appear to have been deposited on a shelf-sea (Jones, 1938, p. lxx).

It therefore appears that the thick Harlech Grit succession lies between areas which underwent shallow-water deposition during the Lower and Middle Cambrian. The stratigraphical and palaeocurrent data tend to suggest that the grits of the Harlech succession were deposited in a NE-SW trough by turbidity currents flowing parallel to the trough axis. At the same time sediments were deposited in shallower water as the marginal and shelf-sea facies to the northwest and southeast. The relationship between the fine-grained flags and the grits of the Harlech succession is not clear. Palaeocurrent observations from the flags are frequently ambiguous and indicate both marginal and longitudinal derivation; the axially deposited material may have been derived marginally and subsequently carried along the length of the basin. Variations in the rate of supply of the gritty detritus resulted in an interdigitation of the marginal and axial facies (Pettijohn, 1949, p. 442) in parts of the geosyncline. Thus, in the Harlech sequence, there is an indication of deposition of sediments derived both along the trough axis and from the trough margins. The inferred palaeogeography is shown in figure 3; this interpretation fits the general concept of the Welsh Lower Palaeozoic Geosyncline in that there is evidence for a

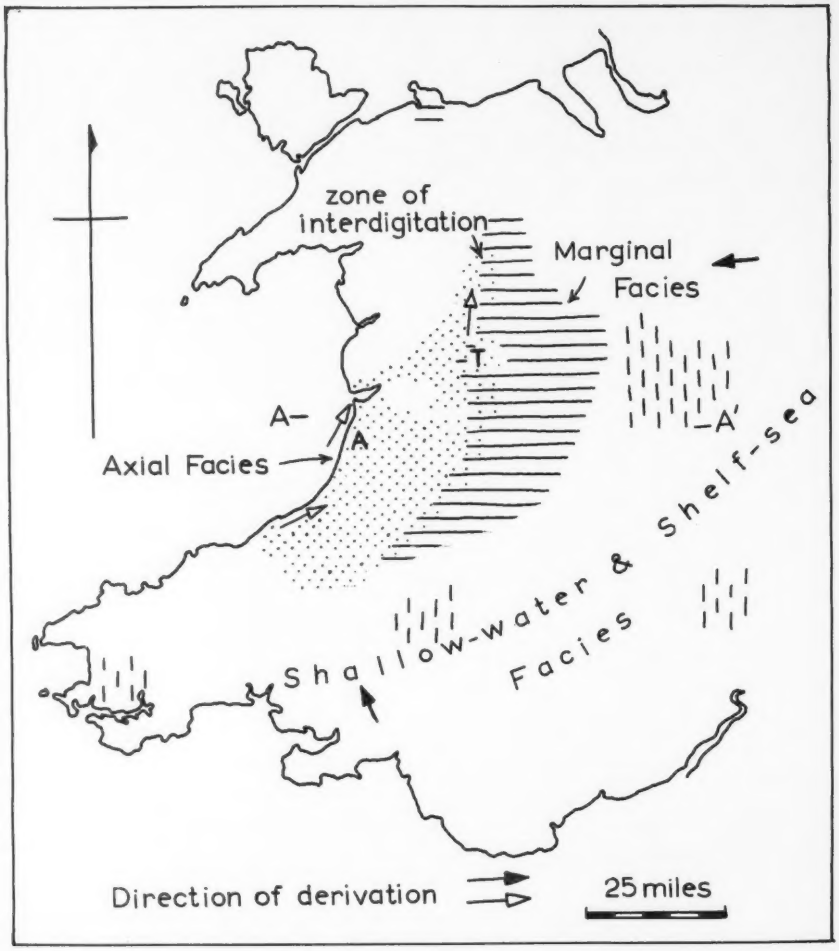

FIG. 4.—Palaeogeography of the Upper Llandovery (based upon Jones, 1938, Bassett, 1955 and Wood and Smith, 1958). A-A′ is the line of section shown in figure 6. Position of Aberystwyth (A) and Talerddig (T) are also marked.

NE-SW landmass lying to the northwest of the Lower and Middle Cambrian sea. The supply of gritty material largely ceased prior to the late Middle Cambrian when dark shales and mudstones appear to have been deposited over much of Wales and Central England. This suggests that the NE-SW basin had become largely infilled by this period.

SILURIAN

A further example of a similar palaeogeographical relationship may be taken from the Upper Llandovery (upper part of Lower Silurian) succession. At the commencement of the Upper Llandovery there was a marine transgression which resulted in the formation of a marginal shelf-sea associated with shallow-water deposits extending over parts

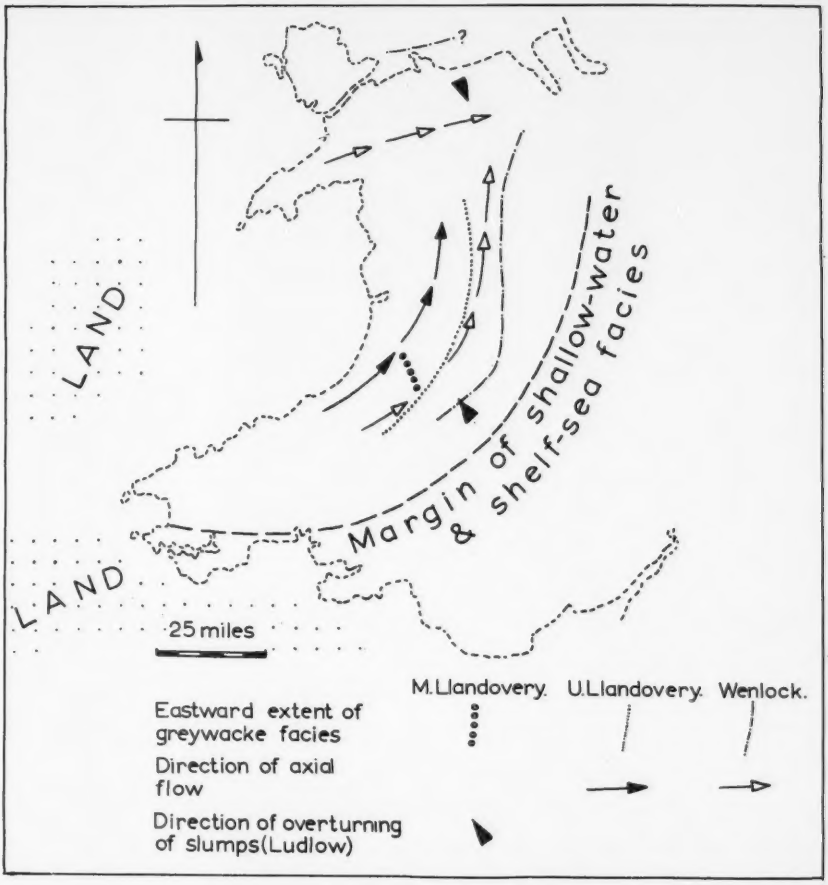

FIG. 5.—Development of the greywacke basin during the Silurian (based upon Jones 1938, Cummins, 1957 and Wood and Smith, 1958) showing the basic palaeogeography, current directions and progressive eastward movement of the margin of the greywackes.

of Wales, the Welsh Borderlands, and Central England (fig. 4). Traced westwards this facies is seen to be replaced by a thicker sequence of dark sediments which includes mudstones, shales, sandstones and greywackes. The greywackes thicken considerably towards the west and also appear at lower horizons in the same direction (Jones, 1938, p. lxxxiv). Palaeocurrent measurements by various authors, Rich (1950), Kuenen (1953), and Wood and Smith (1958) have demonstrated that one group of these greywackes (the Aberystwyth Grits) has been derived from the south-southwest (fig. 4). Bassett (1955) has described another greywacke group from Talerddig (fig. 4); this group also thins eastwards and has been derived from the west or southwest. However, certain of the grey-green mudstones in this sequence (Jones, Bassett, in discussion, Bassett, 1955) may have originated from the southeast, that is, the marginal part of

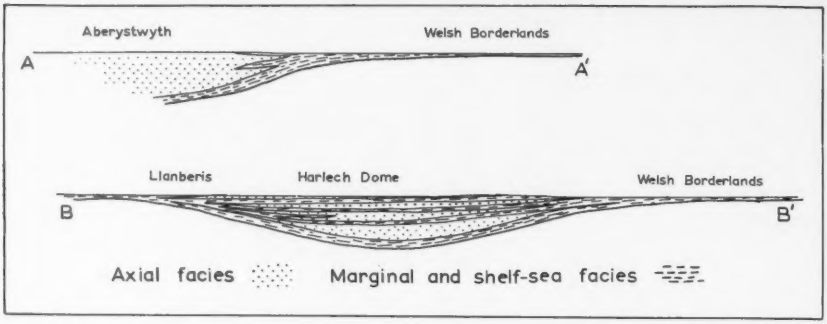

FIG. 6.—Diagrammatic section across the Upper Llandovery (top) and Cambrian basins (lower) showing the differences in interdigitation between the axial greywacke facies and the marginal and shelf-sea facies. The axial facies is largely derived in a direction normal to the section; the marginal facies are derived from the sides of the basins. In the case of the marginal facies of the Harlech succession, evidence of some axial current flow is also present.

the basin. It therefore appears that the distinct sedimentary facies (fig. 4) developed during the Upper Llandovery represented both varying conditions of deposition and differences in the direction of derivation of the sediments. The thick greywacke sequence was deposited in a NNE-SSW trough, largely by turbidity currents; the marginal and shelf-sea facies were probably derived mainly from the southeasterly side of this sedimentary basin (fig. 4). It should be noted that between the axial and marginal facies there is a zone of interdigitation (fig. 4) where the sediments were derived both along the basin and from the basin margins (for example, Talerddig). The Strata Florida Grits (Middle Llandovery), crop out a few miles southeast of Aberystwyth (fig. 5) and appear to have the same lenticular form as the overlying Upper Llandovery greywackes; they thicken rapidly westwards and lie wedged between two successive graptolite zones.

The final example is taken from the Wenlock and Ludlow (Middle and Upper Silurian) successions, deposited during the final phases of the development of this Lower Palaeozoic geosyncline. Cummins (1957) has demonstrated that the Wenlockian greywackes were derived from the south and south-southwest and deposited in two distinct basins (fig. 5). On the southeasterly margins of this same basin, shelf-sea shale/limestone or graptolitic mudstone successions were being deposited. The geosyncline was finally infilled during the Ludlovian. The direction of overturning of sedimentary slumps on the northwestern and southeastern sides of the Ludlow outcrop indicates that there was still a NE-SW topographic depression running across Central and North Wales at this period (fig. 5).

DISCUSSION

Jones (1938, fig. 1, p. lxxvii) has demonstrated that the greywacke facies of the geosyncline extended northeastwards in a diachronous manner during the Silurian (fig. 5). This successive encroachment is related both to the development of the NE-SW trough and to the dominant northeasterly trending currents by which the greywackes were deposited. At the same time, marginal and shelf-sea facies were developed on the sides of the sedimentary basin. Interdigitation of the marginal and axially deposited facies is well developed in the Harlech Dome succession but is less prominent in the Silurian greywackes. The comparatively thin interdigitated zone present in the Upper Llandovery indicates that the basin was better defined at this period than during the Cambrian (fig. 6). This suggests that the sides of the basin were steeper during the later Lower Palaeozoic and tended to confine more closely the axially flowing turbidity currents. This is supported by other lines of evidence.

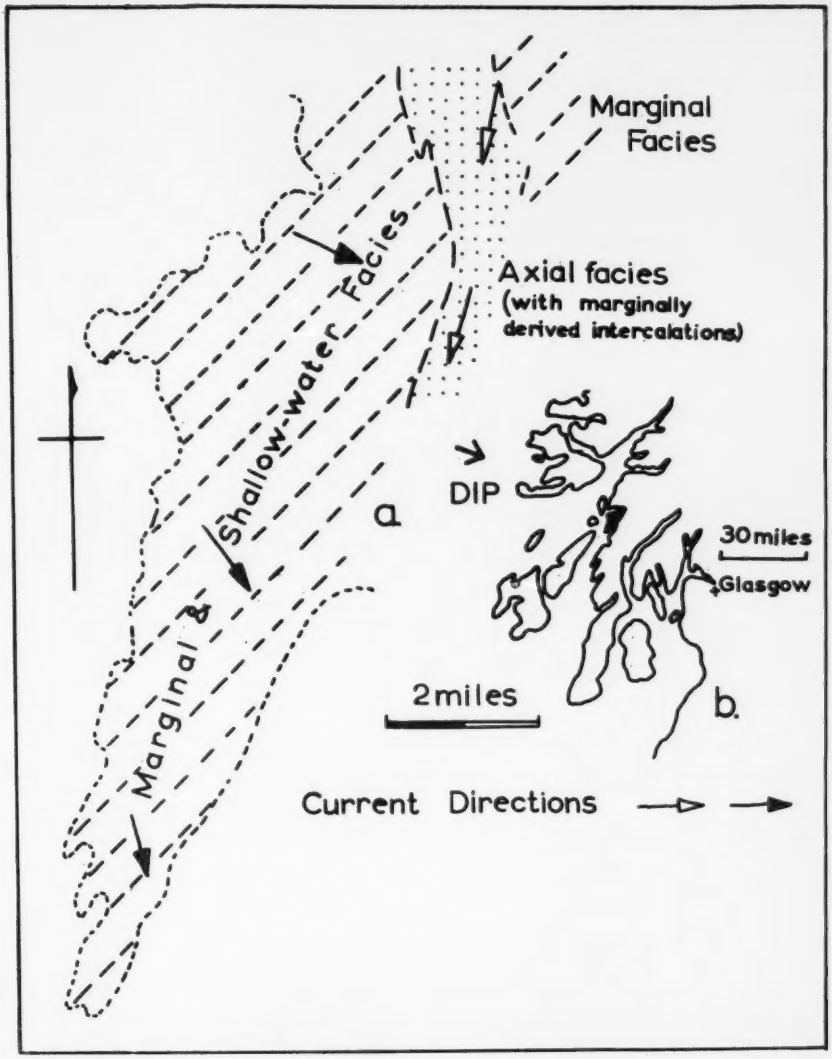

FIG. 7.—a. Axial and marginal sedimentation in the Dalradian of the South-West Highlands of Scotland. b. Locality of the map shown in a.

Graded bedding and other sedimentary structures are better developed on the western side of the Harlech Dome, and this may well indicate the axial zone of the basin where turbidity currents played the major part in deposition. Also, the marked parallelism of the axially orientated flow-structures in the Aberystwyth Grits infers con-

finement of the density currents within a comparatively fixed path.

FURTHER EXAMPLES

The Lower Palaeozoic rocks of Scotland and Ireland also contain abundant greywackes. Although there is little available published data on palaeocurrent directions, it is clear that variations in the direction of derivation of sedimentary material occur; for example, some of the Ordovician greywacke deposits of the Rhinns of Galloway show such variations (Kuenen, 1955, p. 44; Kelling, 1958, figs. 2 and 5).

The Carboniferous Upper Culm Measures of Cornwall and Devon appear to have been derived both along the axis of a sedimentary basin (Butcher, in Kuenen, 1957) and by marginal supply from a land mass to the north of the present outcrop (Prentice, 1956).

In an area of low grade metamorphic rocks in the Southwest Highlands of Scotland, the writer has found evidence for both axial and marginal derivation (Knill, 1959). Part of this area is composed of the Craignish Phyllites which were originally a varied group of shallow-water deposits, derived from the northwest and north-northwest (fig. 7). This group is overlain by the the Crinan Grits which originated, by turbidity currents, from the north-northeast (that is, parallel to the present NNE-SSW tectonic strike). Interbedded with, and also overlying, the Crinan Grits are deposits which have a similar lithology to the upper part of the Craignish Phyllites; these appear to be extensions of the northwesterly derived facies. Sedimentary changes can be recognised in the Craignish Phyllite-type deposits which suggest a variation in conditions of deposition associated with the development of the NNE-SSW basin.

CONCLUSIONS

The recognition of both axially and marginally derived sediments in geosynclinal basins is of stratigraphical importance. Interdigitation of widely contrasted rock types may be related to variations in the direction of derivation of the sedimentary material.

The two facies of sedimentation which have been described, broadly, as "shelly" and "graptolitic" (Jones, 1938, p. lxii) or "shelf" and "basin" now appear to have a new significance. This was recognised by Bailey (1936) when he demonstrated the distinct conditions of deposition of rocks characterised by current and graded bedding, respectively. Not only is the sedimentary environment of these two facies distinct but the derivation may be from entirely different directions or source areas.

Axially deposited material may originate either from the sides or the ends of a geosynclinal basin (Kuenen, 1958). If the sediment is derived from the basin-sides, it must then be transported across the marginal facies in localised channels. Such channels might have a similar form to the submarine canyons of the present day; greywackes, thus, may be marginally derived. The turbidity currents would then turn in direction and flow along the length of the geosyncline. The material derived from the basin-ends might originate from silt-laden rivers, unstable slopes of deltaic fronts or geanticlinal ridges across the tectonic trend of the geosyncline. Kuenen (1958) has emphasised the importance of axial supply in geosynclines but it is clear that, in examples such as the Harlech and Crinan Grits, marginal supply has contributed a large part of the succession.

The tectonic control of the geosynclinal sedimentation described in this paper is clear. The earliest sedimentary basins may well develop parallel to the later mountain chain. As movement continues, the basin will become more closely defined and, thus, the axially and marginally derived sediments become sharply differentiated from one another. The relative rates of development of the geosyncline and the adjacent land mass may well control the nature, amount, and origin of the sediment transported into the basin.

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FIGS. 1–5, SEPTEMBER, 1959

PERMO-PENNSYLVANIAN PALEOGEOGRAPHY OF PART OF OKLAHOMA[1]

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ABSTRACT

Late Pennsylvanian and early Permian rocks, exposed in central Oklaho ma, contain facies changes fossils, pebbles, cross-bedding, and other diagnostic features which permit a paleogeographic map to be drawn for the area. It is recognized that the sea, throughout much of the area, was shallow (zero to perhaps 200 feet deep), and that the sea floor was almost without relief. It is, therefore, obvious that the strand line may have oscillated widely. Nevertheless, agreement among the various bits of evidence is good enough that representative shoreline positions and orientations can be drawn. The sea indicated on the map appears to have withdrawn in, approximately, a northwesterly direction.

INTRODUCTION

The present paper is the synthesis of ideas developed over a period of about 10 years in field work in Oklahoma, arranged approximately as follows: 1947, Hunton Arch and Arbuckle Mountains; 1948, Ouachita and Wichita Mountains; 1950 and 1951, Seminole and adjacent counties; 1952, Ardmore basin; 1955, Osage County; 1956, Noble County and Ardmore basin. The field work carried out in 1951, 1955 and 1956 was done under the auspices of the Oklahoma Geological Survey. Some of the results have been published elsewhere (see references). No report is yet available on Noble County, although one is in preparation, under group authorship.

The geology of part of Noble County will be described briefly, and then results from several parts of Oklahoma will be combined to make a paleogeographic map for late Pennsylvanian and early Permian time.

NOBLE COUNTY

The strata, of Permian age, dip westward at about 30 feet per mile.

Raasch (1946) described the sequence in the northwestern part of the county. He traced the "insect bed," from which were collected more than 5000 insects of 13 orders, four species of Conchostraca (estherians), and two xiphosurans. Carpenter (1947) described many of the insects. Billings (1956) mapped roughly the eastern half of the county, where he found a sequence which is, essentially, alternating shales and thin limestones. Tasch (1958) has been concerned with the Conchostraca. The present author mapped the western half of the county, where the Wellington Formation is exposed. This Permian unit is essentially very fine grained clastics (shales, siltstones, very fine sandstones), with sandstones and a few thin limestones in the section.

A general stratigraphy (compiled from Raasch and Billings) is given in table 1.

In the southern part of the county, the Fallis-Iconium-Garber sequence, alternating sandstones and shales, occupies the same stratigraphic position as the section given in the table. The facies change was mapped by Raasch as taking place along an east-west line through the village of Perry, in the south central part of the county. For this reason, apparently, he mapped the Wellington shoreline as extending east-and-west through Perry. The present author agrees that Perry marks the northern limit of the sandstone facies, but would relocate the shoreline, as is shown below.

The county can be divided into three areas, each characterized by a single facies:

(1) Marine shale and limestone, in the northeastern portion (vertical lines, fig. 1). This facies contains, according to Raasch, the last typically marine fauna in the area. Similar, but older, limestones are found in the section exposed in Osage County, farther east. The Noble County limestones extend southward only a few miles below the latitude of Perry; the Osage County limestones extend almost as far south as the northern edge of Seminole County.

[1] Published with permission of the Director, Oklahoma Geological Survey. Manuscript received December 14, 1958.

TABLE 1.—*Stratigraphic sequence in northern Noble County (after Raasch, 1946, and Billings, 1956).*

	LEONARD
Wellington fm.	Antelope Flats mbr. About 200 ft thick. Mostly sh; thin sss and lss.
	Billings Pool mbr. About 50 ft thick. Sh and ss.
	Midco mbr. About 250 ft thick. Sh, some ss and ls.
	Otoe mbr. About 115 ft thick. Sss and shs.
	Anhydrite sequence. About 190 ft thick. Sh, with thin anhydrite in upper part, lss in lower part.
	Basal sequence. About 15 ft. thick. Ss, some sh.
	WOLFCAMP
Herington ls. About 5 ft thick.	
Enterprise sh (and ss near base). About 57 ft thick.	
Winfield ls. About 5 ft thick.	
Doyle sh (and middle and basal sss). About 80 ft thick.	
Ft. Riley ls. About 5 ft thick.	
Matfield sh. (and upper and basal sss). About 67 ft thick.	
Wreford ls. About 5 ft thick.	

The limestones in northeastern Noble County (Wreford, Fort Riley, Winfield, Herington), are thin (generally less than five feet thick), medium to coarse crystalline limestones. They contain many fossil fragments, algal remains, and clay pebbles. Although sandstones and siltstones are present, the limestones make ledges and control the topography. Bedding planes are rather uneven. Collections of large shell fragments occur locally.

(2) Shales and sandstones, in the southern part of the area (stippling, fig. 1). These sandstones continue, southward, to near the margin of the Arbuckle Mountains. They have been referred to, by many geologists, as the "Fallis-Iconium-Garber delta." Although they were certainly coastal deposits, it is by no means sure that they were deltaic. Cross-bedding studies (summarized in fig. 2) produced three significant directions, 90° from each other, which suggests littoral transportation (Tanner, 1955). The line drawn through the two littoral directions is almost, but not quite, parallel with Raasch's proposed shoreline. If the littoral interpretation is correct, these sands were, at least in part, marine. Since the Osage County limestones extend farther south

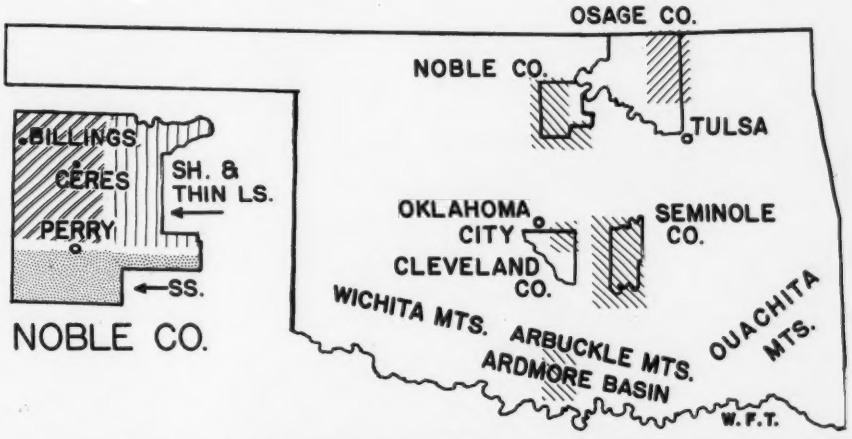

FIG. 1.—Map of Oklahoma showing the outlines of four counties referred to in the text. Shading indicates areas where field work has been carried out. The inset shows Noble County and the three facies found there.

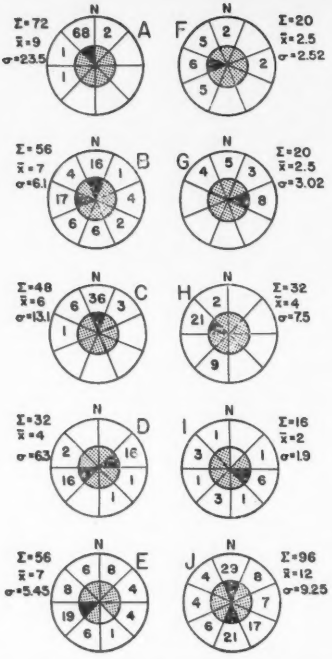

FIG. 2.—Cross-bedding data, not previously published, for the study areas. A, B, and C, summaries of field measurements from Noble County. D, measurements obtained at the Palisades, near Guthrie, between Noble County and Oklahoma City (Garber Sandstone). E. Garber cross-bedding from Cleveland County. F and G, data from early Permian sandstones between Cleveland and Seminole Counties. H and I, pre-Vanoss cross-bedding from the Ardmore basin. J, Vanoss measurements from the Ardmore basin. Numbers outside but adjacent to each compass refer to sum, mean and standard deviation. White octants are "minimum directions," possibly directions toward source areas. Black or dark gray octants have significantly high concentrations of measurements. The test of significance used is: $\bar{x} \pm \sigma$.

than the younger Noble County limestones, these marine sands were spreading farther and farther north with the passage of time.

Limestone and dolomite pebble and cobble (to 3″) conglomerates are fairly common in the sandstones; the pebbles are generally angular to sub-angular. Rubbly limestone (or dolomite) lenses, in red or maroon shales, are of rather limited areal extent, and suggest deposition in temporarily restricted areas; lenses of this type give rise to the "finger-print" soil pattern reported from the air photos (Tanner, 1958). Pelecypod, other shell, and bone fragment conglomerates occur here and there; these fragments show little wear. Plant prints and clay pebbles to about three inches in diameter also occur in the sandstones. Chert or chalcedony pebbles are small and relatively uncommon, in contrast with their abundance in the beds farther south and southeast, where they occur in large numbers and to diameters as great as nine inches. With the exception of the chert, none of the pebbles was transported any great distances.

Ripple marks in the area were produced by currents flowing toward S. 60° W. and N. 20° W., which also are common cross-bedding directions.

Shales in this facies are multi-colored, generally red, maroon, brown, or blue-gray.

Farther south, the rubbly limestone and dolomite lenses do not appear, and the quantity and size of chert pebbles increases. Going southward from Noble County, chert conglomerates begin to appear in numerous exposures in northeastern Cleveland County, about 20 miles southeast of Oklahoma City. From there they increase in size eastward to a maximum in northern Seminole County. The size decrease from Seminole County westward is a decrease with time (younger beds exposed to the west); the decrease from Seminole, Pottawatomie and Cleveland counties, northward, is due to an increase in distance from the source-land, which lay to the south.

(3) Dominantly shales; few thin sandstones, siltstones, limestones and dolomites (diagonal shading, fig. 1). High salinity, at least locally, is indicated by the thin anhydrite beds, the selenite crystals in the shales, and the halite crystal prints, up to two inches on the side, in the limestones. The shales are dominantly gray-green, unlike the shales farther south, and only locally maroon or brown. Fossil wood fragments, silicified bone fragments, clay pebbles, and iron pebbles are found here and there in the area. Beachrock pebbles (Ginsburg, 1953; Tanner, 1956d) up to four inches in diameter occur in the limestones.

Reptile tracks, and possible rain-drop prints, both reported by Raasch (1946), suggest that the area was exposed to the air for at least short periods of time. Mudcrack polygons, quite common in certain thin beds, come in three sizes: one to two inches across, three to four inches across,

and 12 to 18 inches across. They also suggest exposure to the air, although shrinkage cracks have been made in subaqueous muds in the laboratory by the present author. Conchostraca are generally supposed to have been fresh-water forms.

Fossils reported by Raasch from the area include abundant insect prints, fish remains, fern and conifer prints, Conchostraca, and xiphosurans. Algal limestones are common, in beds up to about a foot thick, especially in the townships immediately north of Perry. The algal limestones are laminated, and arched in domes a foot or so across and about an inch high.

Sandstones in the area are, at many places, contorted. Limestones and dolomites are thin, and, unlike the rubbly carbonates in the southern part of the county, continuous. Unlike the limestones to the east, they do not carry a typical marine fauna.

Ripple marks in the sandstones and siltstones were produced by currents flowing toward the N. 30° W. and S. 30° E., an orientation which agrees with the shoreline deduced from cross-bedding data.

The data reviewed in the foregoing paragraphs suggest a nearby shore, and at least occasional exposure to the air. Raasch (1946) accordingly drew a shoreline, in a more-or-less east-west direction, through the site of Perry. The sandstones to the south, for at least 100 miles, however, exhibit primarily littoral type cross-bedding, and therefore the present author prefers to note a shoreline with the following characteristics:

a. East-west or east-north-east, west-south-west-south-west trend, in conformity with cross-bedding and ripple data, and in agreement with the Seminole County shoreline, deduced elsewhere (Tanner, 1955).
b. Position as far south as possible, in conformity with the Seminole County shoreline, and in agreement with the observation that Fallis-Garber cross-bedding is commonly of the littoral type.
c. Wide fluctuations of the strand-line, or local exposures of shoals far at sea, to account for the sub-aerial features observed in Noble County.

CLEVELAND COUNTY AREA

Surface outcrops in Cleveland County, and to the east, are Fallis-Iconium-Garber sandstones and shales, with a few conglomerates here and there. The strike is slightly west of north, and the dip westward at less than one degree. Higher in the section (in the western part of Cleveland County) the Garber Sandstone is succeeded by the Hennessey Shale.

The maximum chert pebble size observed in the area by the present author occurs in T. 10 N., R. 1 W., in northeastern Cleveland County. This is approximately the same latitude as the site of the maximum pebble size in Seminole County. However, the Fallis-Garber sandstones show no evidence of lagoon-and-barrier-island facies, as do the clastics in Seminole County. Instead, many geologists have considered the Fallis-Garber unit as a deltaic deposit.

No marine invertebrates have been reported from these beds, in this area, so far as the present author knows. On the other hand, coal and soot laminae are also unknown. Argument from either of these observations leads precisely nowhere.

The cross-bedding of the sandstones plots in two modes, oriented about 180° apart (fig. 2: E, F, G). In the central part of Cleveland County, Garber cross-bedding is dominantly west or west-south-west. Very faint secondary modes, of dubious reliability, appear to the north and east. In Pottawatomie County, to the east, Fallis cross-bedding is primarily east or east-north-east and west or west-south-west. In the two cases the minimum direction (that is, the direction of least cross-bedding and hence inferred as pointing toward the source) is about S. 25° E. These data are taken as evidence that the Fallis-Garber sandstones, in good part, had a littoral origin.

Baker (1951) examined these two units in considerable detail, obtaining for 214 samples median diameter, sorting, skewness, and heavy mineral counts. For the area studied, he reported median diameter of 0.150 mm, $So = 1.20$, and $Sk = 0.993$. Neither the overall values, nor the local variations, are distinctive enough to be useful in clarifying the environment of deposition.

SEMINOLE COUNTY

The geology of the Seminole County area has been described in detail in a number of publications (Tanner, 1953a, 1953b, 1954a, 1955, 1956b). The beds, of late Pennsylvanian and early Permian age, strike more-or-

less north-south, and dip westward at slightly less than one degree. Typical facies within the area include:

a. Lagoonal or mud flat limestones and dark shales, to the south.
b. Barrier island or bar sandstones and conconglomerates, in the center.
c. Open marine shales, sandstones and limestones to the north.

Conglomerates of the area include limestone, granite, feldspar, biotite, and chert conglomerate pebbles which undoubtedly came from the Hunton Arch-Arbuckle Mountain area to the south, and several varieties of chert which probably were derived from the same area. The less resistant pebbles (such as limestone) coarsen southward toward their source; the more resistant pebbles (such as chert) are coarsest in the middle of the area, and become finer toward both the north and south.

A single conglomerate-bearing channel (in the Vanoss) can be traced linearly in the area; it trends north-north-west. Crossbedding directions are primarily bi-modal (east-north-east, west-south-west), suggesting littoral currents.

Several excellent unconformities permit detailed study of angles and directions of truncation. It is clear that the area has been affected, primarily, by two adjacent areas of uplift (see particularly Tanner, 1956b, pp. 129–130):

a. Toward the east or southeast (and therefore perhaps the Ouachita Mountains); responsible, along with the sinking of the Anadarko basin to the west, for the westward tilting of the entire section, and hence post-depositional.
b. Toward the south (undoubtedly the Hunton Arch-Arbuckle Mountain complex); responsible for the truncation shown along the unconformities, and hence infradepositional.

Since the Hunton Arch-Arbuckle Mountain complex was sufficiently active to (1) furnish materials for coarse conglomerates, and (2) control the truncation of a thick sedimentary section, the complex must have stood as a peninsula, isthmus, or long narrow island during late Pennsylvanian time.

The geometry of the truncations, plus the data reviewed above, suggest a shoreline which trended roughly east-west, or more likely about N. 60° E.

ARBUCKLE MOUNTAINS

The wave-cut Lake Altus erosion surface, so clearly developed in the Wichita Mountains (Tanner, 1954c, 1954d), appears to have as a counterpart the obvious surface of planation which characterizes the Arbuckle Mountains in the vicinity of U. S. 77, and from there west. The Arbuckle Mountain surface in pre-Vanoss, or early Vanoss, in age, because it is overlain in the western part of the mountains, by a thin, undeformed sheet of Vanoss sediments. That it cannot have predated Vanoss time by very much is shown by the fact that all other Pennsylvanian sediments south of the mountains have been highly folded.

At one time, it appears, the East Timbered Hills and the West Timbered Hills must have stood as islands in the late Pennsylvanian sea, while waves rolled across the upturned limestones and dolomites which make up most of the Arbuckle Mountains. During this interval, it is not necessary to draw the Arbuckle Mountains as an isthmus or peninsula, projecting westward into the sea; a couple of islands, beyond a stubby upland, would be more appropriate.

ARDMORE BASIN

Ardmore Basin sediments of Pennsylvanian age correlate with the Pennsylvanian, north of the Arbuckle Mountains, up to about the base of the Vamoosa formation (Alexander, 1953). Vamoosa and Ada equivalents, south of the mountains, appear to be missing. Pennsylvanian rocks older than Vamoosa are thicker south, than north, of the mountains, indicating that subsidence was greater to the south. Similar lithologies were accumulated in the two areas: thick shales and sandstones, with a few thin limestones, and limestone and chert pebble conglomerates.

The Vanoss Formation, which is younger than the Ada, lies unconformably over older Pennsylvanian rocks to the north, west, and south of the Arbuckle Mountains. It is coarsely arkosic to the north and south, but is more typically a shale to the west. The arkosic facies of the Vanoss was supplied

largely from the Tishomingo Granite, exposed in the eastern half of the Arbuckle Mountains, south of Ada and northeast of Ardmore. The igneous rocks exposed in the East Timbered Hills and West Timbered Hills, quite distinct from the Tishomingo Granite, supplied relatively minor amounts of debris to the Vanoss Formation.

Cross-bedding data (fig. 2: H, I, J) obtained from the Pennsylvanian of the Ardmore basin yield the following sets of preferred directions:

a. Vanoss (post-orogenic). North and south directions; inferred to be littoral deposits.
b. Devil's Kitchen (about Savanna or Boggy); Upper Deese (about Wetumka or Wewoka); pre-orogenic. Northwest and southeast, inferred to be littoral; and southwest, inferred to be down-slope deposits.

The Devil's Kitchen also contains oriented plant prints, up to three feet long, with the same orientations; ripple marks, with the same orientations; and imbricated pebbles having intermediate axes dipping toward S. 25° W. The minimum cross-bedding direction, where it is obvious, is about N. 25° E.

Information obtained from orientation data is, therefore, in good agreement with other facts, except for the possible objection that pebbles in the Devil's Kitchen appear to decrease in size toward the south-southeast. In steeply dipping rocks, however, a precise direction cannot be relied upon, because of the restriction of exposures to a narrow band: a flood of detritus from the eastern Arbuckle mountains might very well give rise to a pebble distribution which would have, among others, a south-southeast component.

OSAGE COUNTY

The geology of northeastern Osage County has been described elsewhere (Tanner, 1956c). The beds there include equivalents of the Hilltop and Vamoosa formations of Seminole County, of Missouri and Virgil age. The strike is close to north-south, and the dip westward about half a degree. The strata are chiefly shale, with appreciable amounts of fine-grained sandstone and siltstone, and minor amounts of limestone. The shales are red in the center of the area, and gray-green to the north and south; this is accepted as evidence for a river mouth in the general vicinity.

Oriented data are tri-modal, with two slightly different patterns:

a. In the northern half of the county, two modes are N. 20° E. and S. 20° W. (interpreted as due to littoral currents), and a third mode is N. 70° W. (in the direction of the sea).
b. In the southern half of the county, two modes are N. 20° W. and S. 20° E. (interpreted as due to littoral currents), and a third mode is S. 70° W (in the direction of the sea).

In general, the shoreline ran north-and-south, and the sea lay to the west.

Marine fossils are not common, but do appear here and there in the section. Consequently it is assumed that the shoreline lay to the east of the study area, and, if it was preserved, was contained in rocks which were later removed by erosion.

REGIONAL CONSIDERATIONS

Frank Melton, more than 10 years ago, informally presented four maps showing the geometric relationships between truncating unconformities and shorelines. Diagrammatic versions of these four maps are included in figure 3. The geometric reasoning involved has been outlined in a number of papers (Anderson, 1941; Melton, 1947; Tanner, 1954). The simplest possible case has been reviewed in the present paper, under the heading "Seminole County." This is the case of beds which now strike at, or close to, 90° from their depositional strike, as a result of uplift in two nearby but distinct orogenetic areas. Two such cases can be combined in four different ways to produce the four maps. These (fig. 3) maps represent the following situations:

A. The area of marine deposition has become more synclinal, since deposition (uplift to both north and south, in the figure), and has been tilted up to the east. The major unconformities are shown. The shoreline, at one selected instant, is diagrammed, and the direction of sea advance is shown. This is, of course, a highly generalized pattern, and the actual paleogeography would differ in detail. If this map is viewed so that

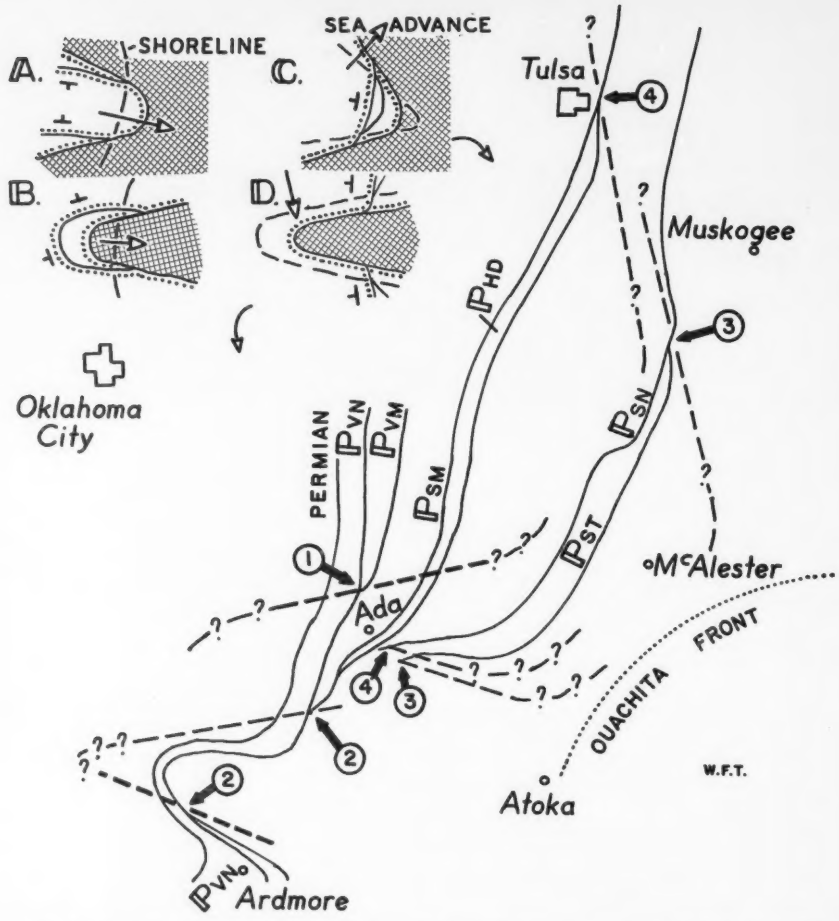

FIG. 3.—Map of east central Oklahoma showing shoreline orientations obtained from truncation data. The simple case (1) is discussed in detail elsewhere (Tanner, 1954b). This case can be combined with its mirror image, in four different ways; the maps which results are sketched in A, B, C, and D, as presented informally by Melton. On each of these maps, a shoreline trend, and a sea advance direction, are shown. Two of these four general cases appear on the map of east central Oklahoma: C, at (3) and at (4); and D, at (2). The shorelines so derived are drawn with dashed lines (contacts traced from Miser, 1954).

the sea-advance arrow points north, this would be a reasonable version of a late Cretaceous or early Tertiary shoreline in the Mississippi embayment area.

B. The area of deposition has become more anticlinal, and has been tilted up to the east. Both shoreline (generalized) and sea advance are shown. If this map is viewed so that the sea-advance arrow points north, this would be a rough approximation of the late Cretaceous or early Tertiary shoreline across Georgia and Alabama. If case B and case A are combined, one obtains a fair paleogeographic map of the

southeastern states, from Arkansas eastward, during late Cretaceous or early Tertiary time.

C. The area of marine deposition has become slightly less synclinal, but has been tilted up to the east. The curvature of the shoreline must be sharper than the curvature of the present outcrops; the overall map pattern is quite obvious and easy to apply. Two examples occur on the map of east central Oklahoma (fig. 3): at the points indicated by the number, 3, where the Senora overlaps the Stuart, and at the points shown by the number, 4, where the Seminole overlaps the Holdenville. The inferred shorelines are shown by dashed lines.

D. The area of deposition has become slightly less anticlinal, but has been tilted up to the east. One example occurs on the map, in south central Oklahoma: at the points indicated by the number, 2, where the Vanoss overlaps older beds. The inferred shoreline is shown by a dashed line.

A fifth sequence of events, obtained when successive uplifts occur in the same area, rather than in adjacent areas, produces no systematic truncation, but instead a line of erosional outliers strung out along the trace of the unconformity. In this instance, the shoreline was roughly parallel with the present strike of the beds (provided no subsequent tilting has taken place in any other direction). Such a situation occurs north of Tulsa, where the Holdenville shale (Memorial shale of Kansas) occurs here and there as outliers along a line slightly east of due north (Oakes, personal communication); a similar instance is diagrammed by Oakes (1940, p. 76), for the Torpedo sandstone of Osage and Washington counties.

The four reconstructions are somewhat more reliable than a shoreline drawn from the simple case (shown by the number, 1, on

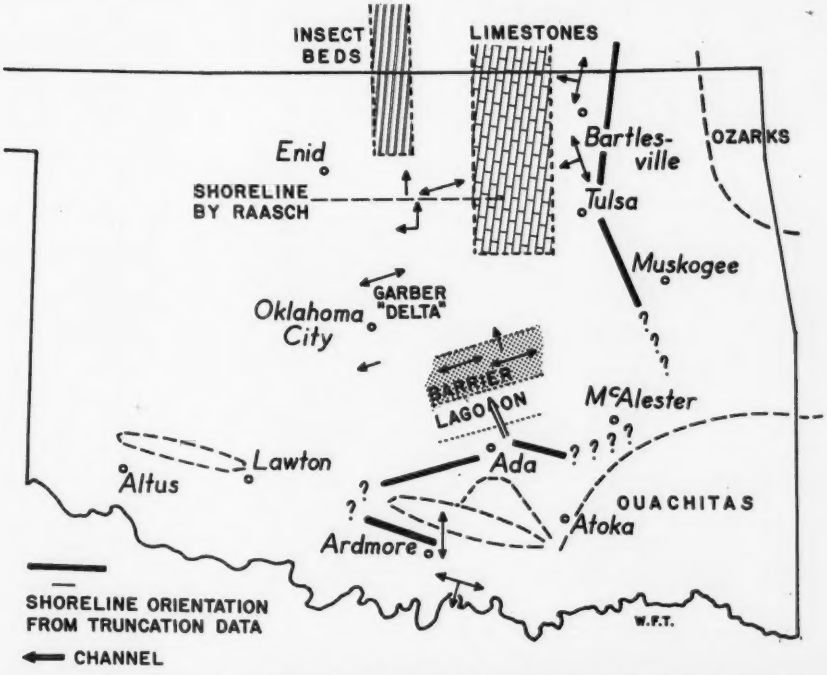

FIG. 4.—Summary of data presented in the text and in previous figures. See text for discussion.

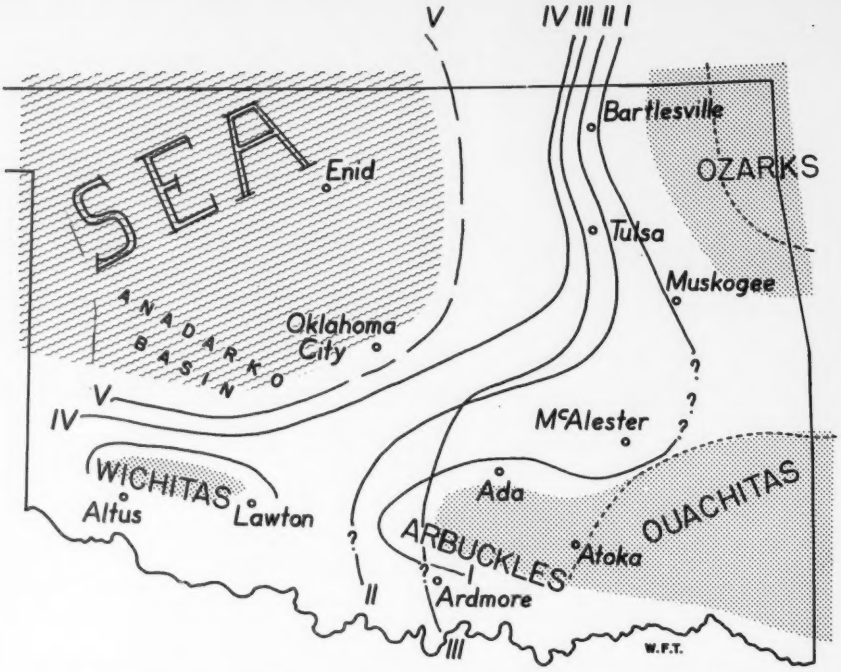

FIG. 5.—Paleogeographic interpretation designed to match the information in figure 4.

the map), because there are two control points instead of one.

The shorelines so obtained are, of course, representative, but not exclusive, of other positions.

SYNTHESIS

The data given in the foregoing sections are summarized in figure 4. On this map, shorelines obtained by use of Melton's diagrams are shown as heavy black lines. Sediment transport directions obtained from crossbedding are shown as slender arrows. A single channel, exposed at the surface and traceable for an appreciable distance, is shown as a double-hafted arrow. The thin marine limestones of north central Oklahoma are indicated by a brick-work pattern, and the insect bed areas of Oklahoma and Kansas are represented by diagonal shading. Stippling marks the inferred barrier of the Seminole County area. The chief mountain ranges of the state are outlined by dashes.

In figure 5 this summary is redrawn in terms of paleogeography. Five different shorelines are presented, with Roman numerals. These range from the oldest, from approximately Stuart time, to the youngest, of post-Wellington time. I through III have been drawn to fit the data presented in the present paper. IV and V have been added to show an eventual sea withdrawal which was more nearly westward than northerly, in keeping with our general understanding of Permian paleogeography.

It is recognized that the late Pennsylvanian-early Permian sea in the area was a shallow shelf sea, with very little bottom relief, and hence that wide fluctuations in shoreline position, and perhaps less spectacular fluctuations in shoreline orientation, undoubtedly took place. The shorelines are thought to be representative, however, and to mark a reasonable progession in time.

The Anadarko Basin, in the western part of the state, does not coincide with the

Permian sea. This should be no suprise to geologists who realize that thickness of section does *not* equate with location of sea; that is, the Anadarko Basin is primarily a structure, rather than a paleogeographic unit, and represents an actively sinking area within a much larger flooded region. To draw Permian shorelines on the basis of the outline of the Anadarko Basin would be to miss completely the lesson of Melton's four type maps, shown in figure 3.

The concept of a shoreline which shifted, with time, seaward in a general northwesterly direction, is not novel. Weaver (1954, p. 96), after studying the rocks of Hughes County, immediately east of Seminole County, wrote that "the irregular lensing of the sandstones . . . , their contained floras and faunas, and the intertonguing of the sandstones and shales in many places are but a few of the characteristics which indicate that the beds were deposited near the shoreline as it gradually shifted to the north and west." This comment was made regarding strata of Missiourian age, slightly older than the beds considered in the present paper. The same idea can be found in older literature.

Four main mountain ranges are shown in figure 5: that all were more-or-less active in late Pennsylvanian to early Permian time is shown by the flood of arkoses surrounding the Wichitas and Arbuckles, the sharp pre-Vanoss truncation of folded Pennsylvanian beds to the south of the Arbuckles, the unconformity system north of the Arbuckles, and the Holdenville-Memorial shale outliers west of the Ozarks. Hence all four are shown on the map. The Wichita and Arbuckle ranges were, perhaps, possessed of greater local relief and more rugged topography than the Ouachitas and Ozarks.

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FIG. 1, SEPTEMBER, 1959

LITHOTOPIC RELATIONSHIPS IN DEEP-WATER TROUGHS[1]

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ABSTRACT

Evidence from modern and ancient deep-water deposits enables the recognition of three lithotopes characteristic of deep-water troughs—a pelagic lithotope, a turbidity current lithotope (with higher- and lower-competency subdivisions) and a slump lithotope. The spatio-temporal relationships of these lithotopes are fundamentally different from those generally obtaining in shallow-water and terrestrial areas. These differences give rise to differences in gross stratigraphic aspect between sequences accumulated in the different areas.

Whilst the Stratigraphic Code is well suited to the description of shallow-water and terrestrial sequences, difficulties are experienced when it is applied to deep-water sequences. An example is cited from near Tamworth, New South Wales. These difficulties are thought to be due to certain limitations in the concepts basic to the Stratigraphic Code, particularly the concept of lithotopic relationships. A re-examination of stratigraphic concepts to test their generality is suggested.

INTRODUCTION

In many ancient and modern deep-water deposits two distinct types of sediments occur. The first of these, consisting of fine lutaceous material, is considered to be a product of pelagic sedimentation (i.e. particle by particle accumulation), and is represented by the oozes and deep-water clays. The second consists of coarser material—silt and sand—sometimes with derived fossils and usually showing good graded bedding. It is produced by the action of turbidity currents (Kuenen and Migliorini, 1950; Ericson and others, 1952) and consists of graywacke suite sediments (in the sense of Packham, 1954).

Ericson and others (1955) record both types in cores from the North Atlantic. Kuenen (1953) recognises the same types in ancient sediments, terming them 'deep-water lutites' and 'graded beds' respectively. These two types are characteristic of many ancient graywacke-bearing sequences.

DEEP-WATER LITHOTOPES

Weller (1958) has recently redefined 'lithotope' as "an area of uniform sedimentary environment." The concept of lithotopes was introduced by Krumbein and Sloss who consider (1951, p. 267) that each distinctive lithology is referable to a distinctive lithotope.

The application of this concept to deep-water deposits leads to the recognition of two broad but distinctive lithotopes which are characteristic of deep-water troughs. These may be termed the deep-water lutite or 'pelagic' lithotope, and the graded bed, turbidite (Kuenen, 1957) or 'turbidity current' lithotope.

Subdivision of lithotopes.—More detailed consideration of the sediments characteristic of deep water would enable subdivision of each of these lithotopes. For the purposes of the present discussion, however, it is only necessary to consider the subdivisions of the turbidity current lithotope.

The sediments deposited in the turbidity current lithotope vary greatly both in grain-size and in the thickness of their sedimentation units. Ericson and others (1952, p. 490) record sand layers "varying in thickness from mere films to beds several meters thick" in modern deep-water sediments. Kuenen (1952), citing Migliorini, has mentioned graded beds up to 10 meters thick.

Kuenen and Carozzi (1953) have pointed out that sedimentation units with the largest grainsizes tend to be the thickest units in any sequence. This is a competency phenomenon. The most competent (and hence largest) turbidity currents produce the thickest and coarsest beds. Variations in competency result in variations in grain-size and thickness of sediment tion units.

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Whilst there will obviously be a continuous gradation in thicknesses from the thinnest to the thickest of units, the thinner units will be more abundant than the thicker both numerically and probably in bulk. The thinner units, which are also the finest, will be grouped together as shale, siltstone, or mudstone units during stratigraphic mapping. Such units may contain a greater or lesser number of coarser (and thicker) units. Once, however, the coarser units become sufficiently thick to be mappable, they may be distinguished as separate arenite stratigraphic units.

It appears practicable, therefore, to recognise two subdivisions of the turbidity current lithotope—a lower competency subdivision and a higher competency subdivision. The former subdivision will be represented by dominantly fine-grained sediment, the latter by coarse-grained sediment.

Minor Lithotopes

Coarse conglomerates are frequently recorded from ancient graywacke-bearing sequences (Natland and Kuenen, 1951; Kuenen and Carozzi, 1953; Crowell, 1957), although they are usually a minor component. These rocks have been attributed to submarine mudflows or landslides, Crowell (1957) having produced weighty evidence of their origin by submarine slumping.

Accordingly a further deep-water lithotope, which may be termed the 'slump' lithotope, may be recognised. Whilst in its broadest sense this lithotope includes all slumping environments and may exhibit sediments of any grainsize, it will be restricted in this discussion to that environment in which coarse conglomerates and pebbly mudstones are deposited. In view of the magnitude of the slumps involved in the production of these lithologies, the sedimentation units so formed may be extremely thick.

RELATIONSHIPS BETWEEN LITHOTOPES

Temporal Relationships

Considered temporally the lithotopes differ markedly in character. The pelagic lithotope is continuous through time from the inception of the deep-water trough until its termination. The slump and turbidity current lithotopes are, however, of intermittent occurrence, and each occurrence is geologically instantaneous. These lithotopes appear in the trough only with the incursion of a slump or a turbidity current, and are replaced by the pelagic lithotope immediately after the slump or current has subsided.

The average time interval between incursions of the turbidity current lithotope is variable, but appears to be large. Kuenen (1953, p. 1046) gives values of from 1000 to 100,000 years between incursions, and Hills and Thomas (1954) have estimated an average of 40,000 years between incursions in an Ordovician sequence in Victoria.

Spatial Relationships

At any particular instant during the history of the deep-water trough, the pelagic lithotope will occupy all areas of the trough, except those where incursions of the turbidity current or slump lithotopes are occurring. The areas occupied by these latter lithotopes are likely to be tongue or fan shaped, and will vary in size depending on the origin and size of the current or slump involved.

In the case of turbidity currents it seems possible that, where these are associated with a nearby submarine canyon, individual currents may be of local extent. Ericson and others (1951) found that sand layers near the mouth of the Hudson Canyon could not, in general, be correlated over distances of 10 miles. However, where a current is initiated on the margin of a trough by a slide unrelated to a canyon, larger areas may be expected. Kuenen (1952) has estimated the area of the deposit from the Grand Banks turbidity current of 1929 at 100,000 sq. miles.

Incursions of the slump lithotope are likely to be of limited extent—of the order of tens of square miles, maximum—as the material involved is relatively immobile once it reaches a relatively flat region.

Spatio-temporal Relationships

Spatio-temporally the various lithotopes will exhibit a repetitive 'interstratified' relationship. The basic lithotope will be the pelagic, which may be said to form a 'background' within which the other lithotopes occur. It is quite possible for incursions of

the other lithotopes to occur with such frequency that the 'background' nature of the pelagic lithotope is not readily apparent. This may be realised if one considers the sedimentary products of these lithotopes, for the volume of turbidite and conglomerate may completely overshadow the deep-water lutites, although the latter represent much longer periods of time.

The frequency of incursions may vary greatly, both in one part of the trough through time, and in different parts of the trough through the same period of time. Thus it may not be possible to match in detail, or even broadly, time-equivalent portions of the sequence in different parts of the trough.

THE EFFECTS OF TECTONISM

Unlike many lithotopes, those under discussion are relatively insensitive to tectonic activity. Tectonism on the margins of the trough may alter the nature and grainsize of the detritus available. It may increase the frequency of incursions, and the relative frequency of higher competency turbidity currents over lower competency types. However it does not destroy the three-fold lithotopic nature of the trough, for this can only be effected by removal of deep water from the trough. This requires extreme tectonism within the trough, and generally terminates its existence.

Thus deep-water lithotopes, as a group, tend to persist for long periods of time. This, added to the frequent development of highlands peripheral to such troughs, often results in the accumulation of very large thicknesses of sediment representative of a restricted number of lithotopes.

STRATIGRAPHIC FEATURES OF DEEP-WATER SEDIMENTARY SEQUENCES

The gross aspect of a sequence deposited in a deep-water trough will depend on the frequency and extent to which the various incursionary lithotopes manifested themselves during sedimentation, and this, in turn, is influenced by tectonism.

Of the several possibilities only one will be discussed, namely that of a trough with quite moderate peripheral tectonism, as an example of a sequence showing features suggestive of this is available. In this case it may be expected that both the higher competency turbidity current lithotope and the slump lithotope will be of relatively minor importance. The bulk of the sequence will consist of deposits characteristic of the pelagic and lower competency turbidity current lithotopes.

These deposits will form an inter-bedded sequence of clays and silts having thin sedimentation units. This will be termed a shale, mudstone, or argillite sequence by the stratigrapher. There will probably be little contrast between the clay and silt units, as the latter will inevitably contain fairly large amounts of the clay characteristic of the former. Silt sedimentation units, whilst traceable for many yards in areas of good outcrop, will ultimately wedge out when traced along the strike.

Within this silt-clay sequence there will occur the deposits of the higher competency turbidity current and slump lithotopes. These will be coarse-grained, and the individual sedimentation units will ultimately wedge out into the surrounding silt-clay lithology when traced along the strike.

In the case of the higher competency turbidity current lithotope the deposits formed will exhibit sedimentation units many times thicker than those characteristic of the silt and clay lithologies. Of these, the deposits from the currents of highest competency and those from a quick succession of currents may be sufficiently prominent to be mappable. These mappable units may be isolated in the silt-clay sequence, or may occur in bunches, each unit being separated by a mappable thickness of the silt-clay lithology.

The deposits of the slump lithotope will probably form thick isolated mappable units within the silt-clay sequence.

The gross aspect of the sequence will be as follows:—a thick sequence of shales, mudstones, or argillites enclosing isolated mappable units of conglomerate and graywacke and/or bunches of discrete graywacke units. All coarse units will be sheet-like and ultimately of restricted lateral extent, wedging out into the surrounding lutite, which occurs between the coarse units. It is as though the lutite were a "stratigraphic matrix" in which the coarser sheet-like units were "set."

This arrangement of lithologies and the great thicknesses of sediment involved are together characteristic of deep-water trough deposits. The arrangement differs markedly from that characteristic of shallow-water and terrestrial deposits, which will now be discussed.

LITHOTOPIC RELATIONSHIPS IN SHALLOW-WATER AND TERRESTRIAL AREAS

In shallow-water and terrestrial areas several lithotopes may adjoin each other spatially at any given time, and all may be persistent through time. The geographic arrangement of lithotopes, their extent, and their permanence is very sensitive to tectonism. An association of lithotopes may migrate laterally through time, either slowly or rapidly, and the migration may be constant in direction or oscillatory. Individual lithotopes or associations of lithotopes may be suddenly obliterated and replaced by other lithtopes. In general, conditions which cause the obliteration of one lithotope tend either to obliterate the whole association, or modify greatly its extent and the spatial relations between its members.

Sedimentary sequences which accumulate in such areas will, in general, be thinner than those of deep-water deposits, and will tend to consist of a succession of several different lithologies without one being dominant. Lithologic units may not be as extensive laterally as is the case with the finer deep-water lithologies, and lateral facies changes often occur. The shape of lithologic units is likely to be tabular, prismatic, or shoestring, rather than sheet-like, and the junctions between lithologic units may be markedly time-transgressive. Where sheet-like units occur, they are likely to be time-transgressive. The result is a three-dimensional arrangement of several different lithologies which, whilst they may intertongue in a complex manner, are each discrete units, and neither enclose the other lithologies, nor are enclosed by them. The Colorado Plateau Mesozoic sequence is a typical example.

Minor exceptions may occur within this arrangement. In such cases one particular lithotope may include small areas of other lithotopes. In general the included lithotopes are transient, and the resultant sedimentary sequence will consist of one dominant lithology containing lenticular masses of other lithologies. This is similar to the situation in deep-water sequences, but the scale is always much reduced, in that the included lithologies are rarely mappable except on very large-scale maps.

APPLICATION OF THE STRATIGRAPHIC CODE TO VARIOUS SEQUENCES

Considering these two types of sequences in the light of the Stratigraphic Code, it will be seen that the Code is admirably suited to the description of the shallow-water and terrestrial sequences. This applies equally to both the Australian Code of Stratigraphic Nomenclature (1956) and the American Code, on which the Australian is based. This appears to result from the basing of the Code on concepts largely derived from consideration of shallow-water and terrestrial sequences.

In applying the Code to deep-water deposits, however, difficulties ensue, as the author has found when attempting to stratigraphically subdivide a late Upper Devonian through Tournaisian deep-water sequence near Tamworth, New South Wales.

This sequence, the Goonoo Goonoo Mudstone[3] (which is being described in detail in another place), is the Lower Burindi Series plus the mudstone stage of the Barraba Series of David (1950, pp. 251–2, 287–291 and Tab. XIV). The mudstone sequence is usually between 10,000 and 15,000 feet thick, and is of deep-water origin, with the exception of the uppermost 2000 feet. A portion of the area occupied by this sequence is shown in figure 1.

The mudstone (unhachured in fig. 1) is in places without included units, but usually encloses several isolated sheet-like graywacke (4, in fig. 1) and polymictic conglomerate units (2 and 3, in fig. 1). These are up to 400 feet thick and wedge out in both directions along the strike. Bunches of graywacke units occur in some areas (1, in fig. 1). The units are generally 50–100 feet thick, and are of limited extent (usually less than five miles) along the strike. These units, of which there are about 3 per 1000 feet of sequence, are completely enclosed by

[3] Pronounced "Gunna G'noo."

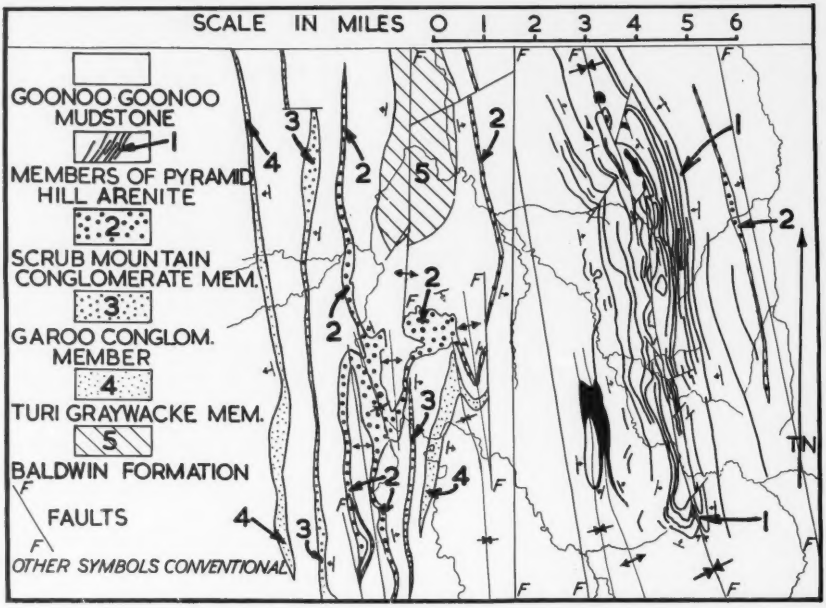

FIG. 1.—Sketch geological map of area 15 miles south of Tamworth, N.S.W.

the mudstones which form the bulk of the sequence.

The sequence is underlain by a mudstone-graywacke sequence containing similar lithologies, but with graywacke units more abundant (5, in fig. 1). This is the Baldwin Formation—the Baldwin Stage of David (1950, p. 251)—and is thought to be a product of a more intense tectonic environment than the overlying Goonoo Goonoo Mudstone. Its junction with the latter is everywhere mappable.

The Goonoo Goonoo Mudstone does not contain any units of sufficient persistence to enable it to be divided into several successive formations without the drawing of boundaries which would be, in part, imaginary, and therefore unmappable. According it has been given formation-status.

The isolated sheet-like but lenticular graywacke and conglomerate units have been termed Members of the Goonoo Goonoo Mudstone (e.g. Turi Graywacke Member, Garoo Conglomerate Member—see fig. 1). A minor difficulty arose at this point, as it was found that at some places within the conglomerate Members the upper parts of the Members were of graywacke and the boundaries of the graywacke-bearing portions were mappable. As there is no provision in the Code for naming subdivisions of Members these variants were not accorded stratigraphic unit status.

In the case of the bunches of graywake units a further difficulty arose. To name each unit individually would have required a multitude of geographic names, due to the large number of units involved and the restricted lateral extent of the individuals. This also would have obscured the essential lithogenetic unity of the units.

It was not possible to treat the bunch of graywacke units as a seperate Formation, as this would have resulted in one Formation containing another Formation. Also it would have necessitated the drawing of an imaginary boundary around the ends of the bunch of graywacke units. As the mudstone between the graywacke units is indentical with and continuous with the mudstone

surrounding the bunch of units, these boundaries would have been, by their nature, unmappable.

For similar reasons it was not possible to treat the bunch of units as a Member within the Goonoo Goonoo Mudstone.

It was decided therefore to depart slightly from the Code and name the bunch of graywacke units as a whole, terming them the "Pyramid Hill Arenite" (1, in fig. 1), which was considered to be a bunch or collection of graywacke Members. The individual Members were distinguished by letter or number suffixes, thus—Pyramid Hill Arenite, Member 1. Each graywacke unit, therefore, is a Member of the Goonoo Goonoo Mudstone, but the bunch as a whole share the same geographic name.

IMPLICATIONS OF THE DESCRIBED EXAMPLE

The difficulties encountered in applying the Stratigraphic Code to the Goonoo Goonoo Mudstone sequence are probably typical of those that might arise during the detailed stratigraphic subdivision of any deep-water sequence. It is here suggested that the Stratigraphic Code may exhibit a certain inflexibility when applied to deep-water depoits in general. This inflexibility may be particularly noticeable when a deep-water sequence is being examined in detail over a large area which has not been greatly disturbed tectonically.

The preceding theoretical discussion suggests that the inflexibility is due to limitations in the concepts basic to the Stratigraphic Code—in particular the concept of lithotopic relationships. The fundamentally different spatio-tempeoral relationships between lithotopes in deep-water troughs, as contrasted with those in shallow-water and terrestrial areas appears to explain adequately the inflexibility of the Code. It is here suggested that the concept of lithotopic relationships basic to the Code be broadened to cover the case of deep-water sediments, and that the Code be modified accordingly to enable easier handling of deep-water sedimentary sequences.

It may be also suggested that the restricted applicability of the concept of lithotopic relationships is part of a larger phenomenon. Almost all our stratigraphic concepts are derived, consciously or unconsciously, from shallow-water or terrestrial sequences. It is well known that deep-water environments exhibit turbidity current activity which differs fundamentally from the traction current activity characteristic of shallow-water and terrestrial environments. It may therefore be expected that concepts which are valid in the latter environments may be invalid when applied to the former environment. The author would therefore suggest a close examination of all stratigraphic concepts to check on their applicability to deep-water environments.

CONCLUSIONS

1.—There are marked differences between the spatio-temporal relationships of lithotopes in deep-water troughs and relationships of lithotopes in shallow-water and terrestrial areas.

2.—These differences give rise to differences in the arrangement of the various lithologies, and differences in gross stratigraphic aspect. In particular, deep-water sequences tend to consist of large thicknesses of a lutite "stratigraphic matrix" enclosing sheet-like coarser units which are large enough to be mappable, but which wedge out into the lutites when traced along the strike.

3.—Difficulties are experienced when applying the Stratigraphic Code to deep-water sequences, and these appear to be due to limitations in the concept of lithotopic relationships basic to the Code, this concept having been derived from shallow-water and terrestrial sequences.

4.—Difficulties may be experienced in matching time-equivalent portions of a deep-water sequence from different parts of a trough, as the incidence of coarse units may vary greatly spatially.

5.—The limitations in the concept of lithotopic relationships may be part of a larger phenomenon. All stratigraphic concepts require checking to test their applicability to deep-water environments.

ACKNOWLEDGEMENTS

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FIGS. 1–15, SEPTEMBER, 1959

MICROFACIES STUDY OF A MIDDLE DEVONIAN BIOHERM, COLUMBUS, INDIANA[1]

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ABSTRACT

The microfacies and statistical study of a Middle Devonian bioherm has revealed that, unlike many similar structures, the investigated one is a composite body in which a transgressive phase of stromatoporoidal growth overlying dolomites has been welded by a horn coral and echinoderm development to a regressive phase of stromatoporoidal growth which is itself overlain by another dolomitic zone. The ecological zones resulting from this strikingly cyclical evolution were found to be, in increasing relative depth order: mat-like stromatoporoids, horn corals and arenaceous foraminifera, echinoderms and ostracods.

INTRODUCTION

The purpose of this investigation was to determine the conditions of establishment, development, and death of a bioherm belonging to the "Coral Zone" of the lower portion of the Jeffersonville Limestone (Middle Devonian), near Columbus, in south central Indiana. A bioherm, as defined in this study, is a lense-like, wave-resistant, structure built mainly by stromatoporoid colonies and horn corals.

LOCATION AND GENERAL DESCRIPTION

The investigated bioherm is located in the Meshberger Stone Company quarry, 2.0 miles northeast of Elizabethtown in Bartholomew County, Indiana, NE $\frac{1}{4}$ sec. 6, T. 8 N., R. 7 E. (fig. 1). The general geology and detailed stratigraphy of the Jeffersonville Formation and adjacent strata in this area have been discussed by Murray (1955). The structure is exposed in the south quarry wall in an area not being actively worked at the present time. This exposure is a part of the Middle Devonian (Onondaga) that extends in a belt from Clark County in southeastern Indiana diagonally across the State to Newton County in northwestern Indiana. Exposures are not frequent other than in quarries and stream cuts due to the thick cover of glacial drift over the area.

The bioherm is 85 feet long with an average thickness of 9 feet (fig. 2); its top and bottom are exposed everywhere, except in two small areas. The basal contact is easy to place within a few inches because of the distinct change in lithology from the saccharoidal, chocolate-brown Geneva Dolomite to the dark blue-gray, biohermal stromatoporoid zone which is usually a little less than 2 feet thick and grades quite rapidly into the overlying light-gray, chalky, coral zone. The latter is soft and porous. It shows a predominance of horn corals and grades into a very "brittle," biostromal limestone composed primarily of mats and fragments of stromatoporoids. Overlying the entire structure is another gray-brown dolo-

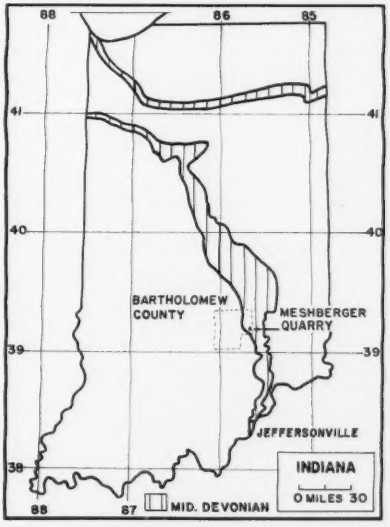

FIG. 1.—Location map.

[1] Manuscript received March 23, 1959.

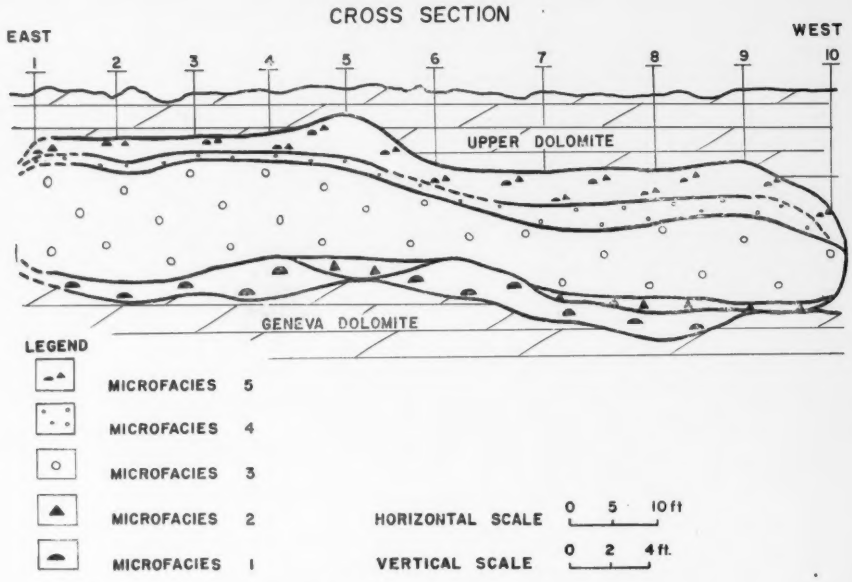

FIG. 2.—General cross-section of the investigated bioherm.

mite several feet thick, followed by a succession of laminated dolomitic limestones and dolomites.

In the field there is little difficulty in distinguishing between the three facies described above, viz., the biohermal stromatoporoid facies, the horn coral facies, and the biostromal stromatoporoid facies. These megafacies correspond to the Microfacies 1, 3, and 5 of this paper. In field observations it is not possible, however, to distinguish Microfacies 2 and 4.

Ten vertical lines were painted about nine feet apart on the wall of the quarry in which the bioherm is located (fig. 2). In each of these sections approximately twelve samples were taken at one foot vertical intervals, starting in the underlying Geneva Dolomite and going as far as possible into the overlying Jeffersonville Dolomite.

DESCRIPTION OF MICROFACIES

The microscopic investigation consisted in the statistical measurement of organic and inorganic parameters in 118 thin sections cut perpendicular to the bedding. On the basis of texture and microfaunal features, the thin sections were found to fit into 7 distinct microfacies:

Geneva Dolomite (fig. 3,a): Fine granular to microgranular gray-brown, massive dolomite with subhedral grains locally changing into rhombohedra. Organic matter and iron oxides emphasize the grain boundaries. "Ghosts" of fossils are present in the upper few feet, but there is a general absence of fossils throughout the formation.

Microfacies 1 (fig. 3,b): Bioconstructed limestone in which mat-like stromatoporoid colonies are separated by irregular zones and pockets of fine-grained bioclastic matrix, consisting primarily of fragments of stromatoporoids, calcareous algae, horn corals, and echinoderm plates. The fragments are irregular and are not sorted. Some recrystallization has taken place in the matrix, and its contact with the stromatoporoids is emphasized by stylolites. Some scattered crystals of pyrite and authigenic quartz are present.

Microfacies 2 (fig. 3,c): Coarse-grained, unsorted,

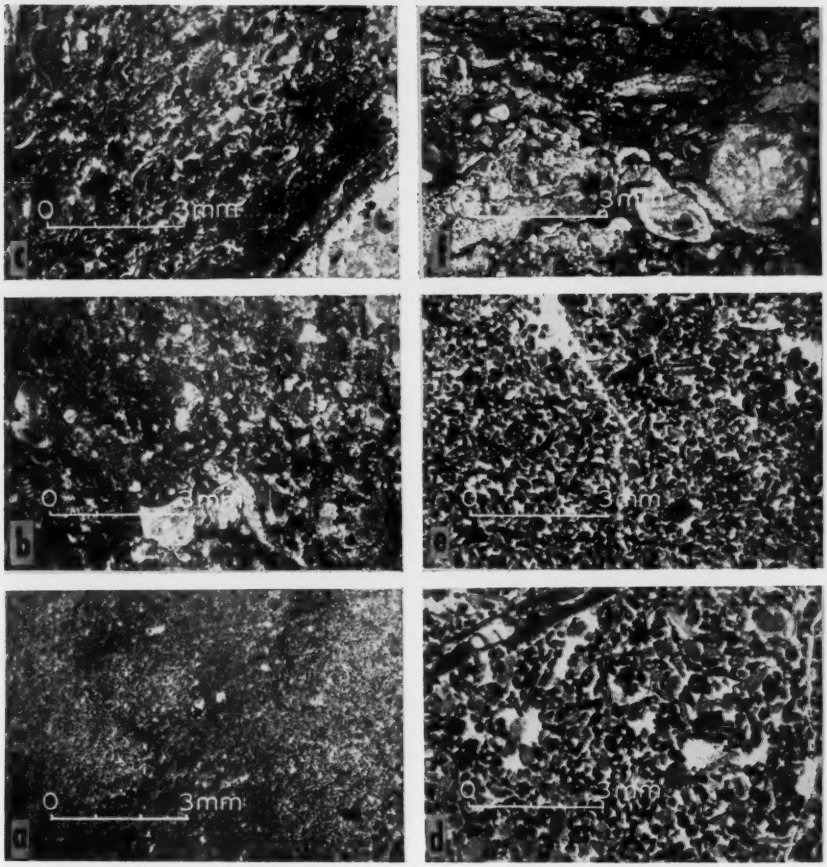

FIG. 3.—Typical microfacies. a: Geneva Dolomite; b: Microfacies 1; c: Microfacies 2; d: Microfacies 3; e: Microfacies 4; f: Microfacies 5.

slightly dolomitic biocalcarenite, with a small amount of well-crystallized calcite cement probably derived from localized recrystallization of a fine-grained matrix. Large and very abundant fragments of stromatoporoids and calcareous algae are associated with grains of dark cryptocrystalline to granular calcite, probably of inorganic origin.

Microfacies 3 (fig. 3,d): Medium-grained, fairly well-sorted calcarenite with well-crystallized calcite cement. It consists of fairly well-rounded pellets and is locally largely recrystallized. There are fragments of echinoderm plates, arenaceous foraminifera (*Textulariidae*), pelecypod and brachiopod tests scattered throughout. Numerous pellets of fine-grained dark calcite, with large fragments of horn corals and stromatoporoids are also present.

Microfacies 4 (fig. 3,e): Fine to medium-grained calcarenite with well-crystallized calcite cement. The organic fragments and pellets are fairly well-rounded and show good sorting. These pellets consist of dark, cryptocrystalline calcite; whereas, the organic fragments are pre-

dominantly echinoderms and ostracods. There is some scattered pyrite and authigenic quartz present.

Microfacies 5 (fig. 3,f): Bioconstructed to coarse bioclastic, slightly dolomitic limestone. Big angular fragments and mats of stromatoporoids and horn corals appear separated by a coarse matrix of unsorted and very angular fragments of stromatoporoids and pelecypod tests. Some relatively large grains of cryptocrystalline calcite are also present with stylolitic contacts. The matrix has an appreciable iron oxide content.

Upper dolomite: Fine-grained, brownish dolomite with "ghosts" of organisms present in some sections. Iron oxide emphasizing crystal boundaries and stylolites is common. The grains are subhedral with a few, scattered, "perfect" rhombohedra.

METHODS AND TECHNIQUES

Field and laboratory techniques were combined in the gathering of data for this study. The most basic of these were the measurements of megafrequencies, microfrequencies, index of clasticity, and calcite-dolomite ratios. All these values were plotted as variation curves alongside the stratigraphic columns (Carozzi, 1958)

Megafrequencies.—To avoid an incorrect interpretation in the thin section analysis due to the large size of the organic fragments in comparison to the small area seen in the thin section, a count was made on the quarry wall of the frequency of occurrence of stromatoporoids and horn corals. A pasteboard frame with an opening of one square foot was placed on the outcrop at the base of the section, and each stromatoporoid or horn coral falling within this area, in whole or in part, was counted as one unit. The frame was moved upward one foot at a time to the top of all but four sections (1, 5, 8, and 10) that were eliminated due to measuring difficulties. The stromatoporoids were generally 3 to 4 inches long and mat-like; however, an occasional specimen was found with a maximum apparent diameter of about 10 inches.

Microfrequencies.—These consisted in the measurement of the total frequency of echinoderm plates, ostracod fragments, and arenaceous foraminifera in 12 randomly distributed fields, each field being of approximately 64 square mm.

Clasticity measurements.—The largest apparent diameter of fragments of colonial organisms, predominantly stromatoporoids, was measured in each slide. These measurements express the capacity of the currents because the stromatoporoids are so massive that their fragmentation is not appreciably governed by internal structures. Usually, clasticity measurements are limited to inorganic fragments (Carozzi, 1958).

Calcite-dolomite ratios.—Samples taken from the field were ground to a very fine powder by mortar and pestle for a period of three minutes. They were then placed in an electric grinder for a period of forty-five seconds. Each sample was then shaken through a 0.088 mm standard sieve, placed in a bottle, sealed, and labelled. Sealing the bottle was a precaution taken against the sample's becoming hydrated prior to its being X-rayed and therefore causing erroneous intensities of the calcite and dolomite peaks. When ready for X-raying, the prepared powder was tapped lightly into a slot (1 by 2 cm) cut into an aluminum holder and backed by a small glass plate. The powder was compacted; great care was taken not to orient the particles. The sample was then inserted into an X RD-3 General Electric diffractometer fitted with a copper X-ray tube. The percentages of calcite and dolomite were calculated on a basis of the sum of their respective peaks being equal to 100 percent of the carbonate sample. From this, simple percentages of each variable could be computed.

The errors introduced in the X-ray analysis through humidity, preferred orientation, and differential hardness of samples are thought to be insignificant.

DESCRIPTION OF SECTIONS

In figures 4 through 14, the lithologic symbols used to represent the microfacies are identical to those of figure 2.

Section 1 (fig. 4).—Section 1 demarcates the eastern edge of the bioherm. This is the most poorly preserved section due to the concentration of ground-water effects at this point. The calcite-dolomite percentages show that the bioherm itself is composed entirely of calcium carbonate and that it is

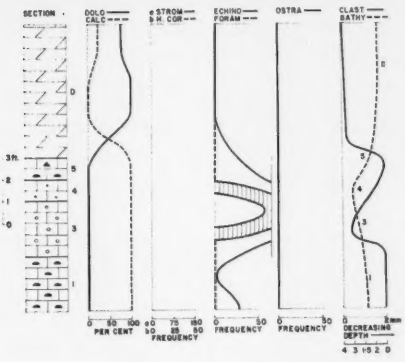

FIG. 4.—Section 1.

overlain by nearly pure dolomite. Stromatoporoids and horn corals were not counted in this section due to the irregularity of the surface produced by ground-water action. Echinoderm frequency reaches a maximum in the upper portion of Microfacies 3 and the bottom of Microfacies 4. Ostracods and benthonic, arenaceous foraminifera are present but are in numbers too small to permit any reliable counting. Organic clasticity varies from infinity in Microfacies 1 and 5 (infinity being here 2 mm and larger sizes in bioconstructed zones) to 0.32 mm in Microfacies 3.

Section 2 (*fig. 5*).—Section 2 contains all of the microfacies except no. 2. The calcite-dolomite percentages indicate that the bioherm is limestone surrounded by dolomite. Stromatoporoids are prominent in Microfacies 1 but decrease sharply in Microfacies 3 and 4. In direct contrast, horn corals, though almost entirely missing in Microfacies 1, are prevalent in Microfacies 3, declining again in Microfacies 4 and 5, with the return of the stromatoporoids. Echinoderms develop a peak in Microfacies 4 paralleling a smaller one of the arenaceous foraminifera. The latter develop their strongest peak in Microfacies 3 at the same time as the ostracods. Minimum organic clasticity is approached in Microfacies 4 with a value of 0.16 mm as opposed to infinite values in Microfacies 1 and 5.

Section 3 (*fig. 6*).—Section 3 is used as a key section throughout this study due to the excellent exposure it displays. Calcite-dolomite percentages are the same as in section 2. The stromatoporoid frequency reaches a peak in Microfacies 1, declines steadily through Microfacies 3, and rises again to a secondary peak in Microfacies 5. Horn corals, however, are poorly represented in Microfacies 1 and do not come into prominence until the stromatoporoids show a marked decline in their frequency. Echinoderms persist in all microfacies though not strongly represented until Microfacies 3; they show a partial parallel variation with the ostracods. Arenaceous foraminifera show peak values at the base of Microfacies 3, varying directly with the horn corals. The ostracods show two peak values, one in

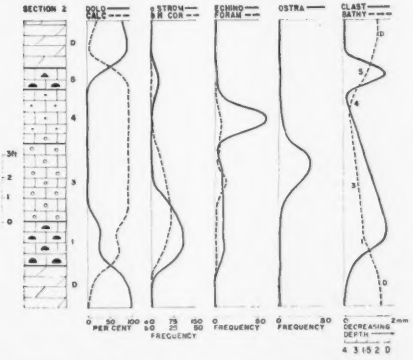

FIG. 5.—Section 2.

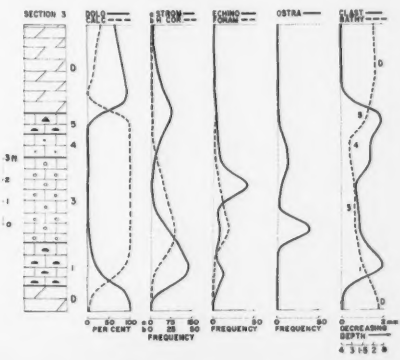

FIG. 6.—Section 3.

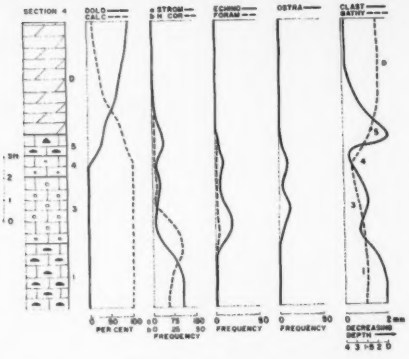

FIG. 7.—Section 4.

Microfacies 3 with the arenaceous foraminifera and horn corals and another smaller one in Microfacies 4. Organic clasticity varies from infinity in Microfacies 1 and 5 to 0.68 mm in Microfacies 3.

Section 4 (*fig. 7*).—In this section, the calcite-dolomite percentages are the same as in the preceding sections. The stromatoporoid frequency varies inversely to horn coral frequency, with the former dominating in microfacies 1 and 5 and the latter dominating in Microfacies 3. Echinoderms exhibit two peak values, one in Microfacies 3 and another in Microfacies 4. The arenaceous foraminifera are present but in numbers too small to be interpreted. The ostracod frequency shows parallel variation with that of the echinoderms. Organic clasticity varies from infinity in Microfacies 1 and 5 to 0.14 mm in Microfacies 4.

Section 5 (*fig. 8*).—This section is the first one described so far that contains representatives of all the microfacies. Calcite-dolomite percentages are the same as in the previous sections, with the exception of a slight increase in the dolomite percentage at the top of Microfacies 5. The frequencies of stromatoporoids and horn corals were not taken due to sampling difficulties. The echinoderms show a first peak in Microfacies 1 followed by a decrease in frequency in Microfacies 2; then, a long maximal zone appears to culminate in Microfacies 4. The arenaceous foraminifera are present in small numbers throughout the section but show no clear peak values. The ostracods show a broad peak in Microfacies 4, similar to the one shown by the frequency of the echinoderms at that level. Organic clasticity varies from infinity in Microfacies 1 and 5 to 0.94 mm in Microfacies 3.

Section 6 (*fig. 9*).—Microfacies 2 and 4 are not present in this section. The calcite-dolomite percentages are the same as in the previous sections. The stromatoporoids show high frequency values in Microfacies 5. Horn corals have their greatest development in Microfacies 3 and show a secondary peak in Microfacies 5. The echinoderms display a first peak that roughly corresponds with that of the stromatoporoids in Microfacies 1 and a second peak at the top of Microfacies 3.

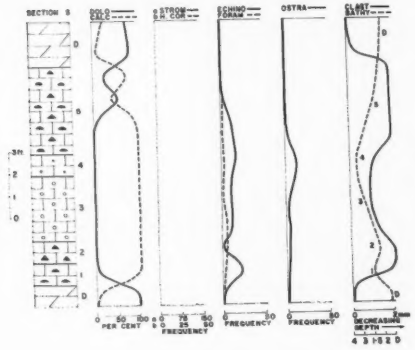

FIG. 8.—Section 5.

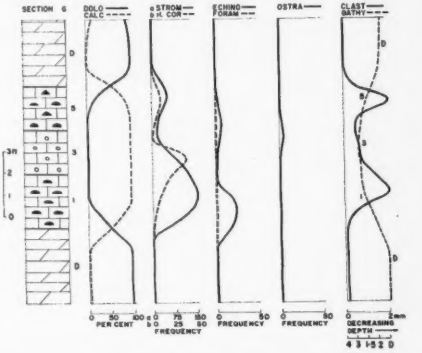

FIG. 9.—Section 6.

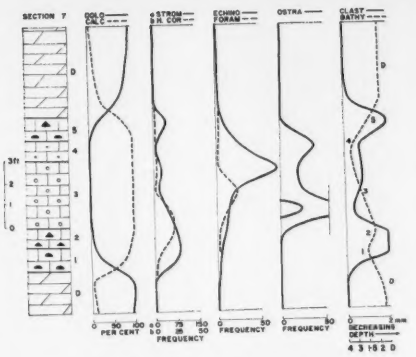

FIG. 10.—Section 7.

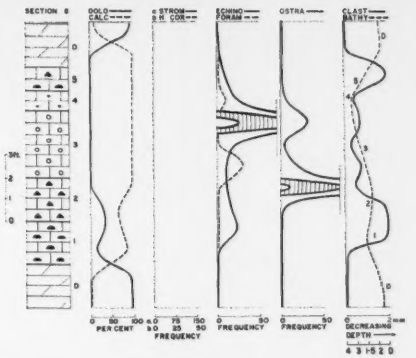

FIG. 11.—Section 8.

The arenaceous foraminifera are constant throughout the section at minimum values. The ostracods are essentially constant except for a small increase of frequency toward the top of Microfacies 3 in association with the echinoderms. Organic clasticity varies from infinity in Microfacies 1 and 5 to 0.51 mm in Microfacies 4.

Section 7 (*fig. 10*).—This section displays all the microfacies, and the calcite-dolomite percentages are the same as in the preceding sections. The stromatoporoid frequency is irregular but has essentially the same trend in relation to the horn corals as has been described previously. This relationship is a general decline of the stromatoporoids with a corresponding increase in horn corals. The echinoderms show a marked peak at the top of Microfacies 3 following those shown by the arenaceous foraminifera and the ostracods. The latter reveal another but less important peak in Microfacies 4. On the whole, the ostracods seem to show a much higher frequency of occurrence on the western side of the bioherm than on the eastern side. Organic clasticity varies from infinity in Microfacies 1 and 5 to 0.28 mm in Microfacies 3.

Section 8 (*fig. 11*).—This section also contains all of the microfacies. The calcite-dolomite percentages are the same as in the previous sections. The frequency of the stromatoporoids and horn corals is not included due to sampling difficulties. The echinoderms are fairly constant but reach an extremely high peak in the upper portion of Microfacies 3 in correspondence to a similar, though less intense one of the ostracods. The arenaceous foraminifera show an almost parallel variation with the ostracods. The latter reach two distinct peak values, one very pronounced in Microfacies 2 and another less pronounced in the upper portion of Microfacies 3. Organic clasticity varies from infinity in Microfacies 1 and 5 to 0.32 mm. in Microfacies 3.

Section 9 (*fig. 12*).—This section contains all microfacies except Microfacies 1. The calcite-dolomite percentages are the same as in the previous sections. The stromatoporoid frequency is variable but very definitely decreases with increasing numbers of horn corals. The echinoderms show a marked increase in Microfacies 5 reaching there a peak

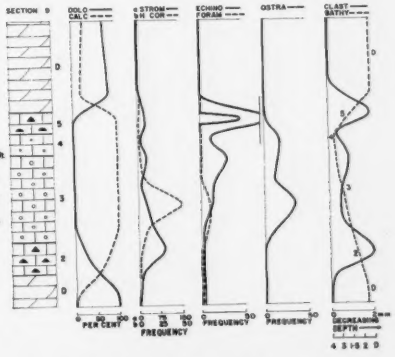

FIG. 12.—Section 9.

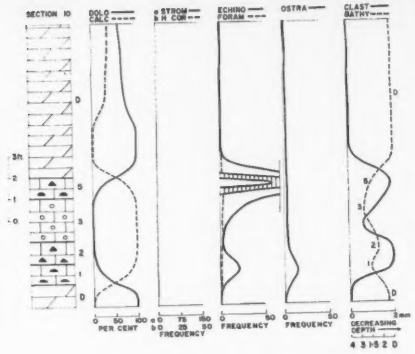

FIG. 13.—Section 10.

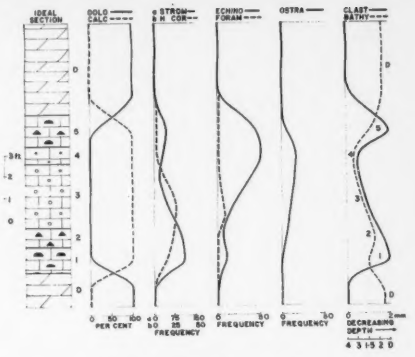

FIG. 14.—Ideal section.

value preceded by a smaller one at the top of Microfacies 3. The arenaceous foraminifera show a small peak at approximately the same level in Microfacies 3 as the horn corals and the ostracods; the ostracods display a slight increase of frequency in Microfacies 4. Organic clasticity varies from infinity in Microfacies 2 and 5 to 0.42 mm in Microfacies 4.

Section 10 (fig. 13).—Section 10 represents the western end of the bioherm and the pinching together of the microfacies. This end of the structure is much more apparent in the field than the eastern one. The calcite-dolomite percentages are the same as in the preceding sections. Sampling difficulties prevented the determination of the stromatoporoid and horn coral frequencies. The echinoderms display two peaks, a small one in Microfacies 1 in correspondence with the ostracods and a much larger one in Microfacies 5. The arenaceous foraminifera do not vary throughout the section. Organic clasticity varies from infinity in Microfacies 1 and 5 to 0.48 mm in Microfacies 3.

IDEAL SECTION AND RELATIVE BATHYMETRICAL INTERPRETATION

The five microfacies that constitute the bioherm should be considered not as separate, unrelated strata, but instead as units that reflect the environmental changes of the sea in which it developed.

The ideal microfacies sequence of the bioherm can be worked out by combining the ten investigated sections, as in figure 14; from the bottom to the top it is as follows:

1. bioconstructed limestone formed by matlike stromatoporoid colonies
2. coarse-grained, unsorted biocalcarenite formed mainly by fragments of stromatoporoids
3. medium-grained, well-sorted calcarenite with abundant horn corals and arenaceous foraminifera
4. fine to medium-grained, fairly sorted calcarenite with abundant echinoderms and ostracods
5. bioconstructed to coarse, unsorted bioclastic limestone with associated stromatoporoids and horn corals.

The calcite-dolomite percentages clearly show that the bioherm is entirely calcium carbonate, surrounded by dolomite. By the use of the X-ray techniques previously described, it is possible to place within a few inches the contact between the bioherm and the dolomite at the base of the structure. The top contact is a little more difficult to place precisely, due to a small percentage of dolomite (about 15 percent) in the upper portion of Microfacies 5.

Stromatoporoids exhibit two major frequency peaks. The lower peak in Microfacies 1 is much more strongly expressed than the upper peak in Microfacies 5. The former corresponds to a well-developed biohermal zone which developed immediately on top of the Geneva Dolomite when normal marine conditions were restored by deepening of the sea. The second peak is a thinner,

biostromal zone formed during the shallowing phase which led to the return of the dolomite facies in the Jeffersonville. The frequency curve of the horn corals shows a low value in Microfacies 1 followed by a gradual increase through Microfacies 2 towards a maximum in Microfacies 3. From this point upward a steady decline is apparent. The horn corals developed best in deeper water between the two stromatoporoidal zones and particularly in the medium-grained calcarenite immediately overlying the biohermal zone. They decreased rapidly in the overlying fine to medium-grained calcarenite (Microfacies 4) which corresponds to a still deeper environment.

The frequency curve of the arenaceous foraminifera shows a variation parallel with that of the horn corals. As the latter, the foraminifera were eliminated due to deepening of the sea.

The echinoderm frequency varies from a minimum in Microfacies 1 through an almost continuous increase to a maximum in Microfacies 4. It is clear that they developed immediately after the decrease of the horn corals. The ostracods are at minimum values in Microfacies 1 and also continuously increase in frequency upwards, reaching a maximum in coincidence with that of the echinoderms. The parallelism of variation between benthonic echinoderms and pelagic ostracods is not frequently reported and can be explained here by the fact that Microfacies 4 represents the maximum deepening of the sequence, these conditions almost stopped the development of the horn corals and arenaceous foraminifera, favored the expansion of the echinoderms and at the same time allowed free access to the open sea and migration of the pelagic ostracods. The curve of the organic clasticity shows peak values of 2 mm in Microfacies 1 and 5, which as previously mentioned are considered to be infinite, due to the large size of stromatoporoids corresponding to biohermal and biostromal conditions. The clasticity decreases gradually through the non-sorted calcarenites of Microfacies 2 (1.70 mm) into the sorted calcarenites of Microfacies 3 (0.70 mm) and 4, reaching the minimum values in the latter with 0.50 mm. Naturally, clasticity values are at zero in both overlying and underlying dolomites. By combining the indications given by the organic components and the clasticity, we can classify the described microfacies according to relative increasing bathymetrical depth:

1. dolomites underlying and overlying the bioherm
2. unsorted biocalcarenite formed by stromatoporoid fragments (Microfacies 2)
3. bioconstructed stromatoporoidal limestones (Microfacies 1 and 5)
4. medium-grained, well-sorted calcarenites with horn corals and arenaceous foraminifera (Microfacies 3)
5. fine to medium-grained, fairly well-sorted calcarenites with echinoderms and ostracods (Microfacies 4).

The relative bathymetrical curve shows that after the stromatoporoids were established on top of the Geneva Dolomite as a result of increasing depth, a shallowing followed due to the rapid growth of the colonies which made them intersect wave action (Microfacies 2). After this destructive phase, depth of water increased gradually in Microfacies 3 with predominance of horn corals and arenaceous foraminifera and in Microfacies 4 with predominance of echinoderms and ostracods, reaching its maximum value in the latter. A rapid decrease of depth took place leading to the return of the stromatoporoids but without redevelopment of Microfacies 3, some horn corals are found, however, associated with the biostromal zone of Microfacies 5. The overlying dolomites mark the return of conditions similar to those of the basal Geneva Dolomite. The relative bathymetrical curves drawn for each section (fig. 4 to 13) show variations around the above described generalized evolution resulting from the local absence of one microfacies. It is interesting to consider the asymmetrical character of the depth variation during the evolution of the bioherm corresponding to a slow transgression followed by a rapid regression, starting and ending in dolomitic facies.

During the transgressive phase, we have noticed the succession of three ecological zones. These are listed in order of increasing depth:

1. mat-like stromatoporoids
2. horn corals and arenaceous foraminifera
3. echinoderms and ostracods.

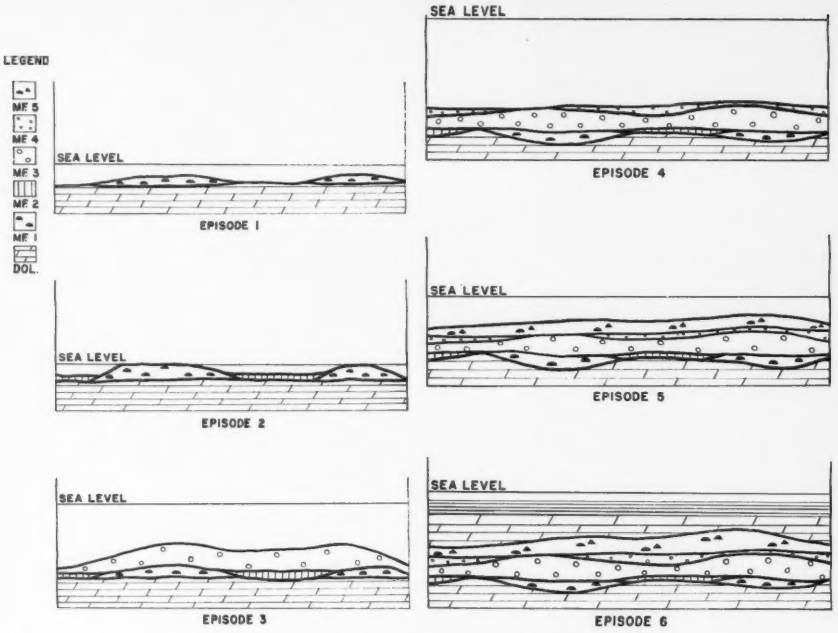

FIG. 15.—General bathymetrical interpretation (episodes 1 to 6).

A similar sequence of zones has been described by Lecompte (1958) for Devonian bioherms in Belgium. These too, may be listed in order of increasing depth:

1. massive stromatoporoids
2. mat-like stromatoporoids
3. colonial corals
4. brachiopod-bearing shales
5. barren shales
6. goniatite-bearing shales with light pelagic fauna.

In the case investigated here, the massive stromatoporoids were not observed, and the zone of echinoderms and ostracods is equivalent to zones 4 to 6 of the Belgian occurrences.

GENERAL CONCLUSIONS

By integrating all the information derived from the field and the microscopic investigations, one could formulate the following bathymetrical interpretation of the development of the bioherm.

1. The Geneva Dolomite and the Jeffersonville Dolomite which respectively underlie and overlie the bioherm appear to be of primary origin, resulting from the direct precipitation of a high calcium-magnesium sediment from lagoonal, iron-rich waters. Had secondary dolomitization taken place, it is almost certain that any rock with the high porosity of the bioconstructed material would have been dolomitized very easily.
2. Microfacies 1 (fig. 15, episode 1) represents a stage of environment in which the sea was extremely shallow but a little deeper and clearer than the water from which the Geneva Dolomite was apparently precipitated. At this time the stromatoporoids and the calcareous algae began to populate the area. They attached themselves to the soft mud bottom and formed the first bioconstructed accumulations.

3. Microfacies 2 (fig. 15, episode 2) is a product of reworking of Microfacies 1, accumulated in low areas between the stromatoporoid mounds which had grown up to the wave action zone.
4. Microfacies 3 (fig. 15, episode 3) represents the beginning of the deepening phase during which the environment was ideal for the establishment of the horn corals which quickly took over the realm of the stromatoporoids and reduced them to a small percentage of their former number.
5. Microfacies 4 (fig. 15, episode 4) represents the time of greatest submergence. The organic clasticity is at a minimum recorded for the average of any of the microfacies, indicating the weakest bottom currents. The general decline in horn coral remains suggests that their lower limit of depth had been reached, leading to the development of the echinoderms and the introduction of pelagic ostracods. It is probable that the weight of the bioconstructed zone, at that time, was sufficient to cause differential compaction of the underlying Geneva Dolomite, causing the undulating surface upon which it now rests.
6. Microfacies 5 (fig. 15, episode 5) corresponds to a rapid shallowing during which stromatoporoids and a few horn corals formed a biostromal to bioclastic layer again in the zone of wave-action.
7. The upper dolomite (fig. 15, episode 6) is similar to the Geneva Dolomite in appearance though it is not as dark. This deposit indicates the return of very shallow and lagoonal waters.
8. Several feet of laminated dolomitic limestones and dolomites with numerous mud-cracked layers overlie the upper dolomite. These were not studied in detail, but it is believed that they represent the final stages of the drying up of the lagoonal areas.
9. Unlike most bioherms which result from a continuous phase of organic growth ending under the effect of a change in environment, the investigated structure, in spite of its apparent homogeneous look in the field, is actually a composite bioherm in which a transgressive phase of organic growth has been welded to a regressive one.

ACKNOWLEDGMENTS

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FIGS. 1–5, SEPTEMBER, 1959

STORM SEDIMENTS ON A PACIFIC ATOLL[1]

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ABSTRACT

Deposits resulting from a single violent typhoon that struck a Pacific atoll consist of (1) coarse gravel ridges on the reef flat, (2) new or augmented beach ridges of gravel, (3) gravel sheets, ranging from one-half to three feet in thickness, across two-thirds of some islets, and (4) sediment redeposited on the lagoon sides of many islets. These storm deposits differ greatly in texture and structure from those formed in corresponding places under normal environmental conditions. In addition, storm sediments of the lagoon floor are relatively coarser at any particular depth than those formed under other conditions. Both storm and normal types are recognized in sections that expose earlier islet deposits.

INTRODUCTION

Jaluit Atoll, in the southern Marshall Islands, is located at approximately 6° N. and 169½° E. in the central Pacific Ocean. It consists of a roughly lozenge-shaped peripheral reef, 84 nautical miles in length and up to 1650 feet in width (fig. 1) surrounding a lagoon. The lagoon has maximum dimensions of 33 and 25 nautical miles. Along the reef are many small islands, four of which were examined in detail during the present study.

A severe typhoon (Ophelia) hit Jaluit Atoll on January 7, 1958. On some islets nearly all trees were uprooted and swept away, soils and fine sediments were scoured out and removed, and layers of coral rubble were spread over the surface. Extensive geomorphic changes likewise resulted; many land forms were destroyed and new features were developed. Because of the intensity of this storm a survey was made in late April and May, 1958, to determine details of physical and biological change. The project was sponsored by the Pacific Science Board of the U. S. National Academy of Sciences—National Research Council, in cooperation with the Office of Naval Research and the U. S. Trust Territory Administration. The field party consisted of seven scientists representing various fields of study under the leadership of David I. Blumenstock of the U. S. Weather Bureau.

Deposits resulting from typhoon Ophelia are very different from those developed under normal conditions of sedimentation as represented by the partly removed and partly buried deposits of the former islet surface. Storm sediments of the type examined are deposited only occasionally in any one area. The preceding severe storm to strike Jaluit was more than 50 years ago. Elsewhere in the Pacific, the average occurrence at any particular place of typhoons of similar intensity is believed to be not more often than once every 30 to 50 years. The sedimentary records of such typhoons, however, are recognized in many test pits and trenches on various islets, and indicate that the deposits of violent storms probably comprise a considerable proportion of the total sediment in some places.

The purpose of this paper is to record characteristics of the deposits from a devastating storm[2] and to indicate differences between these and deposits formed under normal depositional conditions. Both types apparently contribute in varying proportions to the total record of sedimentation.

Acknowledgement is made to J. I. Tracey, Jr., and to Curt Teichert of the U. S. Geological Survey for helpful suggestions concerning the problem and for criticism of the manuscript.

DEVELOPMENT AND GROWTH OF ISLETS

The peripheral reef of Jaluit Atoll, like

[1] Publication authorized by the Director, U. S. Geological Survey. Manuscript received May 18, 1959.

[2] An earlier report on observations of the effects of a violent storm (cyclone) on a coral island is by F. W. Moorehouse (1936) who was residing on Low Isles, Great Barrier Reef of Australia, at the time. Subsequently the effects of this cyclone were evaluated and mapped by Fairbridge and Teichert (1948) on the basis of new aerial photographs

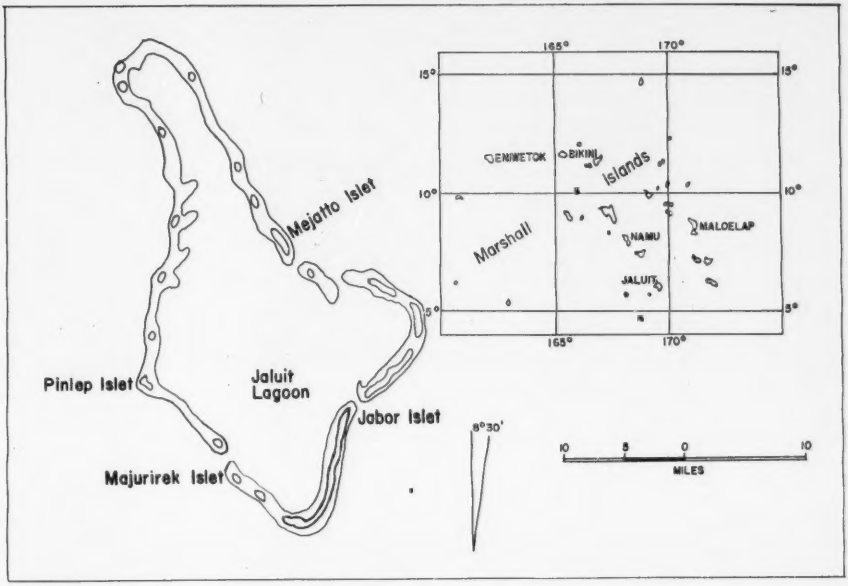

FIG. 1.—Location map of Jaluit Atoll.

that of other atolls, is composed of rigid, wave-resistant skeletons of sessile corals and coralline algae, with clastic particles or unbroken shells and skeletons of benthonic organisms which partly or entirely fill cracks and interstices. In contrast, rocks that rest upon these reefs and that normally form the islets rising above them, are very different structurally and texturally. Such rocks consist entirely of detritus from the reef, ranging from sand to boulder size, and of the skeletons of organisms, all cemented to varying degrees. These rocks may or may not show well-developed stratification. Bedding is poorly defined and inconspicuous where coarse material has been laid down in broad sheets or as mounds; it is prominent and has primary lagoonward dips where normal sorting processes of a beach have been responsible for its development.

The forming of islands upon the peripheral reefs of atolls is generally attributed to the accumulation of detritus at a surface irregularity or nucleus, during a particular stand of sea level. Should sea level rise suddenly and appreciably, a probable result would be rapid upward growth of reef-forming organisms so that even the former island area might be covered with the new reef rock; should sea level go down the island doubtless would be destroyed by subaereal erosion. With a relatively constant position of sea level, however, an island may develop within certain limits as a result of geological processes operating under two types of conditions: (1) the normal condition of day by day deposition and erosion resulting from waves, tides, longshore currents, and other regular controls; (2) the occasional great storms which act violently, and abruptly modify the normal environment.

In order to interpret correctly the history of any particular islet on an atoll, the processes operating under each of the two conditions cited above must be understood and appraised, and criteria must be established for recognizing the deposits formed by each. Clearly, most islets are comprised of deposits representing both normal and storm conditions, but the proportions attributable to

each on any particular islet differ widely.

In general, the deposits of normal sedimentation on an islet consist of sand and fine gravel with good sorting and well-developed cross-stratification. Constant reworking of present beaches by waves and tides tends to remove the very fine materials (below sand size) and to separate fine gravel and sand into layers. Because permanent accumulation of sediments is largely in the lee of the islets, such sediments form a lagoonward extension of beach deposits and therefore islets normally build forward in that direction.

Islet sediments deposited during major storms, in contrast to those deposited in normal weather, consist dominantly of widespread gravel of pebble to boulder size that seems to be the product of rapid mass movement. The gravel forms ridges along the windward shores and blanket deposits across large parts of islet interiors. It may also form temporary ridges on the reef flat. In general, these storm deposits are characterized by relatively poor sorting, rude stratification, and by fair to good imbrication of pebbles and cobbles. Removal of sandsize and smaller particles through winnowing action is widespread.

The past history of certain islets on Jaluit Atoll can be deduced in part through examination of sections both in natural exposures and in artificial wells and trenches. On the islet of Mejatto, for example, exposures in a section (fig. 3D) 1½ miles south of the northern end show consolidated, cross-stratified calcium carbonate sandstone and conglomerate, with laminae dipping lagoonward. This deposit is about 150 feet from the present seaward margin of the islet although it must have developed on a former margin of the lagoon. Such once-buried remnants of earlier beaches indicate the extent to which beach sands in this area have migrated lagoonward across the reef during the history of the islet.

On Jabor Islet, in a narrow part about 1¼ miles from the north end, a trench dug across the land almost to low tide level illustrates that here, on the other hand, little or no beach sand is present (fig. 2A). In this section, typical reef rock at the bottom of the trench is overlain by 3 feet of brown, well-cemented conglomerate which was apparently formed under storm conditions early in the development of this islet. White, poorly consolidated but otherwise similar gravel, which overlies the conglomerate, apparently had a similar origin at some later date. Thus, in this part of the islet there is no evidence that rocks were formed by normal beach accumulation of fine sediments.

Cross-sections through Mejatto Islet (figs. 2D and 3A and B) illustrate variations in the relative contributions of the two types of deposits (normal and storm) preserved from early stages of islet development. These variations are similar to those described from Jabor.

MODIFICATIONS OF ISLET STRATA RESULTING FROM TYPHOON

A principal objective of the present study has been to determine and record the effects of typhoon Ophelia on the geomorphic and structural features of islets on Jaluit Atoll. This has been accomplished by examining the sediments in detail, measuring the strata, and plotting in cross-section available data for two islets—Jabor and Mejatto—known to have been especially hard hit and awash during the storm. Because effects of the typhoon on these islets include both accretion and removal of sediments, an attempt has been made to indicate the distribution and extent of these changes.

Sedimentary deposits attributed to typhoon Ophelia that have added to the bulk of Jabor and Mejatto Islets consist of unweathered or very slightly weathered gravel ranging from pebble to boulder size, with very little interstitial sand or other fine particles. The gravel consists in part of material torn loose from the reef front. Such fresh sediment is readily recognized by its uniformly white color, in contrast to older gravels that are gray or brown as a result either of algal covering or of weathering in a soil zone. Deposits of flat gravel commonly contain imbricated pebbles and cobbles, the surfaces of which dip in the direction from which the storm waters advanced.

Based upon their geomorphic position, gravel accretions resulting from the typhoon may be divided into three classes. These are: (1) gravel ridges that locally form bars on the seaward parts of the reef; (2) shore ridges; and, (3) gravel sheets or blanket deposits.

Reef ridges of gravel are especially well-

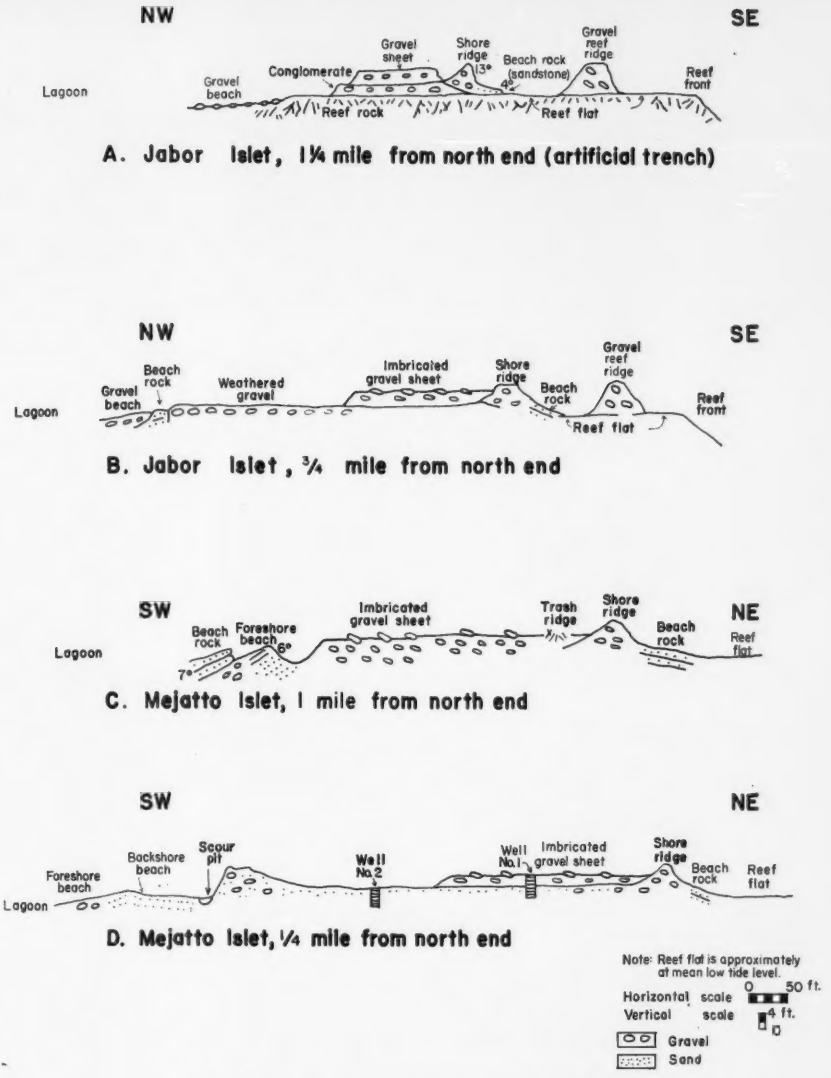

FIG. 2.—Sections across Jabor and Mejatto Islets.

developed on the reef flat seaward of Jabor Islet (fig. 2A and B), where for most of the length of this islet they are 8 feet high and 45 to 60 feet wide; the ridges are not so well-developed on the reef flat seaward of Mejatto Islet. In both places they contain abundant blocks and boulders from 1 to 5 feet in diameter, many of which were recently derived from the reef front as shown by their fresh little-worn surfaces and by the

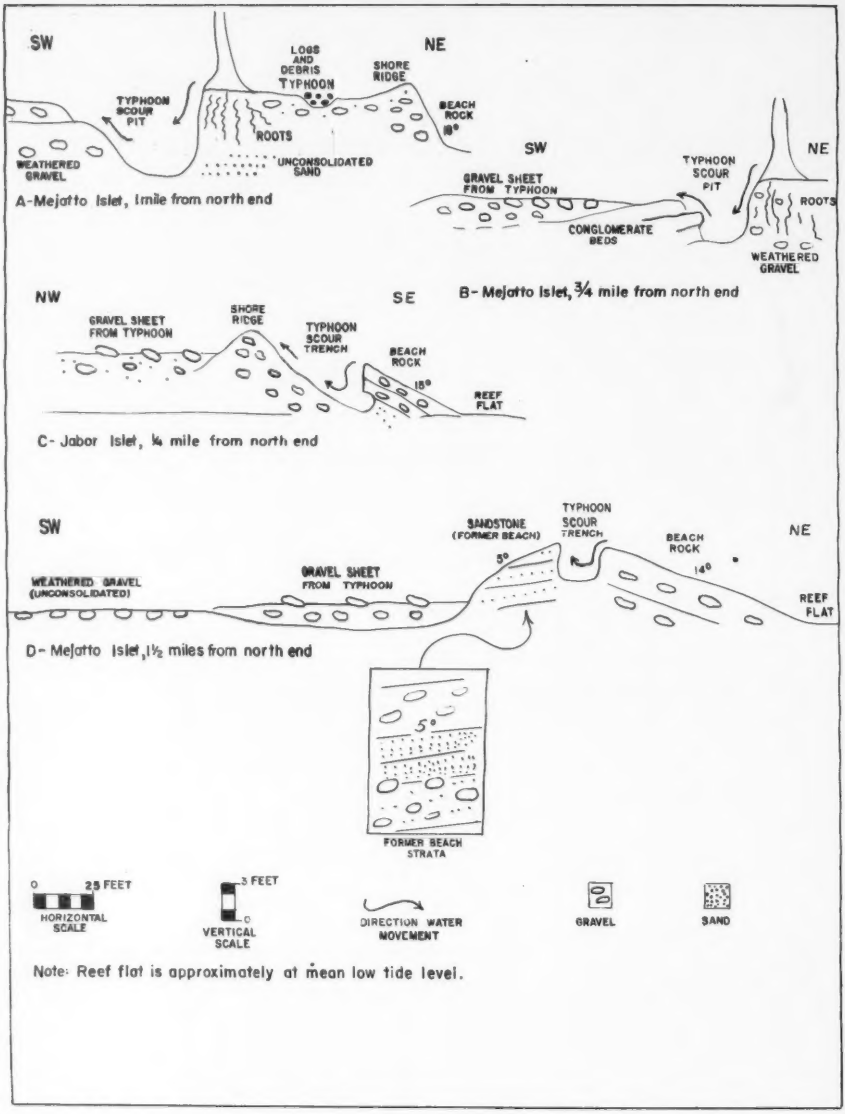

FIG. 3.—Scour trenches and pits and gravel deposits formed by typhoon Ophelia on Jabor and Mejatto Islets.

types of coral represented. Sections across the reef ridge on Jabor Islet show conspicuous imbrication of large rock slabs which dip seaward. This imbrication indicates that the slabs had been transported landward. At the time of examination, which was three months after typhoon Ophelia, these reef ridges had already migrated a considerable distance landward, according to observations of J. B. McKenzie and others who had been on the ground during that interval. There seems little reason to doubt that normal wave processes will eventually carry the gravel farther back and add it to the seaward deposits of the islet.

Shore ridges on the seaward sides of Jabor and Mejatto Islets are like the reef ridges in all esssential respects except location. They represent a climax in the accumulation and piling up of coarse debris. These shore ridges are higher and contain larger boulders than other geomorphic forms on the islets and they indicate the maximum concentration of storm-deposited debris along the islet front. Structurally, also, they seem similar to the reef ridges and they probably are enlarged by material from these ridges during landward migration.

The most significant quantitative additions to the islets during typhoon Ophelia are the blanket deposits of gravel, here termed gravel sheets. These are thin layers of white, relatively unweathered gravel deposited on large parts of those islets that were inundated by storm waters. They seem to have been spread out and deposited in the manner of river flood or glacial outwash deposits. Gravel sheets on Mejatto Islet (figs. 2C and D, 3A and B) begin on the seaward side directly lagoonward of the shore ridge, or of the scour trenches and pits (plunge holes) as layers of loose gravel a few inches thick. In places, they extend two-thirds to three-fourths of the distance across the islet, where they end abruptly in a ledge or nearly vertical drop of two or three feet along a sinuous front. Texturally they are distinctive because of the absence of sand or other fine matrix. Structurally they form a single bed or layer, but individual slabs commonly show imbrication.

Gravel sheets on the islets differ considerably in particle size from place to place as shown on Mejatto and Jabor Islets, but in general the particles are considerably finer than in gravel deposits of the shore ridges and the beach ridges. Gravel sheets seem to have been derived from at least three sources: (1) the outer reef area; (2) earlier shore ridges; and, (3) reworking and redistribution of gravel from older sheets, with a winnowing away of soils and fine materials. It was not possible during the present study to determine the relative contributions from each of these sources. A significant observation, however, is that enough gravel was introduced from outside the islet to appreciably raise the general island level and to deposit a new stratum of gravel as a record.

Although much sediment, nearly all coarse, was deposited on islets by floodwaters during typhoon Ophelia, notable erosion also resulted from these waters. Evidence of such erosion is especially conspicuous in areas on the islets that were apparently occupied by relatively weak sediment adjacent to resistant surfaces. Scour trenches several feet deep were cut into unconsolidated sand deposits on both Jabor and Mejatto Islets landward from, and parallel to, the upturned edges of beds of resistant beach rock that dip toward the sea (fig. 3C and D). On Mejatto Islet there are many examples of plunge holes or scour pits in weak deposits of sand, extending to the lee of areas where the sediment is tightly bound by the root systems of trees (fig. 3A and B). One of these plunge holes is 6 feet deep. Thus, with the advance of water from northeast to southwest, the unprotected areas southwest of obstructions on the seaward sides of the islets were selectively scoured.

Erosion was also considerable in areas that bordered the lagoon shores of islets, especially between the margins of newly deposited gravel sheets and resistant beach rocks of the lagoon edge. In the unconsolidated sand of such areas as found on Mejatto (fig. 2C and D), the rushing water apparently concentrated in channels to scour out large plunge holes which have subsequently formed tidal pools.

A large amount of fine sediment, especially of sand size, obviously was removed from the islets of Jabor and Mejatto by flood waters of typhoon Ophelia. This fine sediment was winnowed out of the sandy gravel of the island and was largely stripped from

former sand areas. At the time of the investigation, the only significant sand areas that were exposed on these islets were in the bottoms and along the sides of deep scour trenches and in undercut areas around trees and beach rock. Some of the removed sand apparently then formed bars that extended out into the lagoon at various places; some of it also was deposited as submarine offshore bars (fig. 4), developed by waves. The great bulk of the sand, however, presumably has been carried into the lagoon, whose floor is now considered to have been raised appreciably.

REMOVAL OF FINE SEDIMENTS FROM ISLETS

Sand and soil are almost completely absent from Jabor and Mejatto Islets, both of which were awash during typhoon Ophelia. That such sediments formerly constituted significant parts of the surfaces of these islets is indicated by the prevalence of sand and soil on the unflooded northern end of Mejatto Islet and on nearby islets such as Pinlep and Majurirek which were not flooded. Not only has the former beach sand been removed, but also sand and soil seem to be thoroughly winnowed out from among the cobbles and pebbles which now form a gravel sheet across much of these islets.

The destination of fine materials removed from the islets by storm waves has been determined by investigations of two types. First, a study has been made of sediments from the lagoon shore outward to determine the present position of various size grades of sediments. Second, using aerial photographs a comparison has been made of the offshore geomorphic features of 1944 with those of 1958.

The lagoon area off Jabor Islet was selected for examination of bottom sediments because it represents a place in which storm intensity and islet flooding reached a maximum. Analyses were made of samples collected at 100- to 200-foot intervals outward from the shore to depths of 60 feet (fig. 4). These show only coral gravel from the shore line outward to 180 feet, poorly sorted limesand from 180 to 1000 feet, and an accumulation of leaf-like and branch-like coral fragments (*Montipora* sp. and *Acropora* sp.) beyond 1000 feet. This distribution is largely attributed to typhoon Ophelia. The lack of sand on the lagoon beach in this area contrasts with the abundant sand on lagoon beaches developed by normal wave and tidal action. Poor sorting of the offshore sands and the lack of progressive decrease in median grain size with depth and distance offshore, suggest rapid deposition and consequent mixing. In these respects they differ considerably from offshore sediments reported from Kapingamarangi Atoll (McKee, Chronic, and Leopold, 1959, figs. 5, 7, 8, and 9) which were deposited under conditions of normal sedimentation.

Additional features of the lagoon floor off Jabor Islet that may be attributed to typhoon Ophelia are an offshore sand bar or ridge which lies parallel to and 400 feet out from the shore, and a large accumulation of pandanus trees resting on the sand floor at the 50-foot depth, directly beyond a steep drop-off 600 feet from shore (fig. 4). Thus, the storm has left a record offshore consisting of poorly sorted fine sediments and land-derived plants which, if buried and preserved, will appear very different from normal offshore deposits.

Lagoon sediments at Pinlep and Majurirek Islets, where the typhoon effects were also great, but where flooding did not occur as on Jabor, were also studied. Relatively little fine sediment was removed from these islets. Nevertheless, the offshore lagoon sands are poorly sorted (fig. 4) as on Jabor, and samples seem to indicate a considerable amount of mixing as far offshore as 700 feet, and at depths as great as 15 to 25 feet. In contrast, lagoon beaches on these islands were formed of sand, the analyses of which show good to fair sorting similar to that of beaches developed under normal conditions of reworking by waves and tides.

Constituents of offshore sand in the Jaluit lagoon are shown by sample counts to consist largely of broken and worn pieces of coral, although mollusk shell fragments are also very common in all size grades. The tests of foraminifers, relatively uniform in size, make up more than 50 percent of the particles in the coarse-grain size, but they are scarce in other size grades. Other contributions, including sea urchin spines and parts of *Halimedas*, are quantitatively unimportant. Comparison of these sediments with those accumulated in similar locations

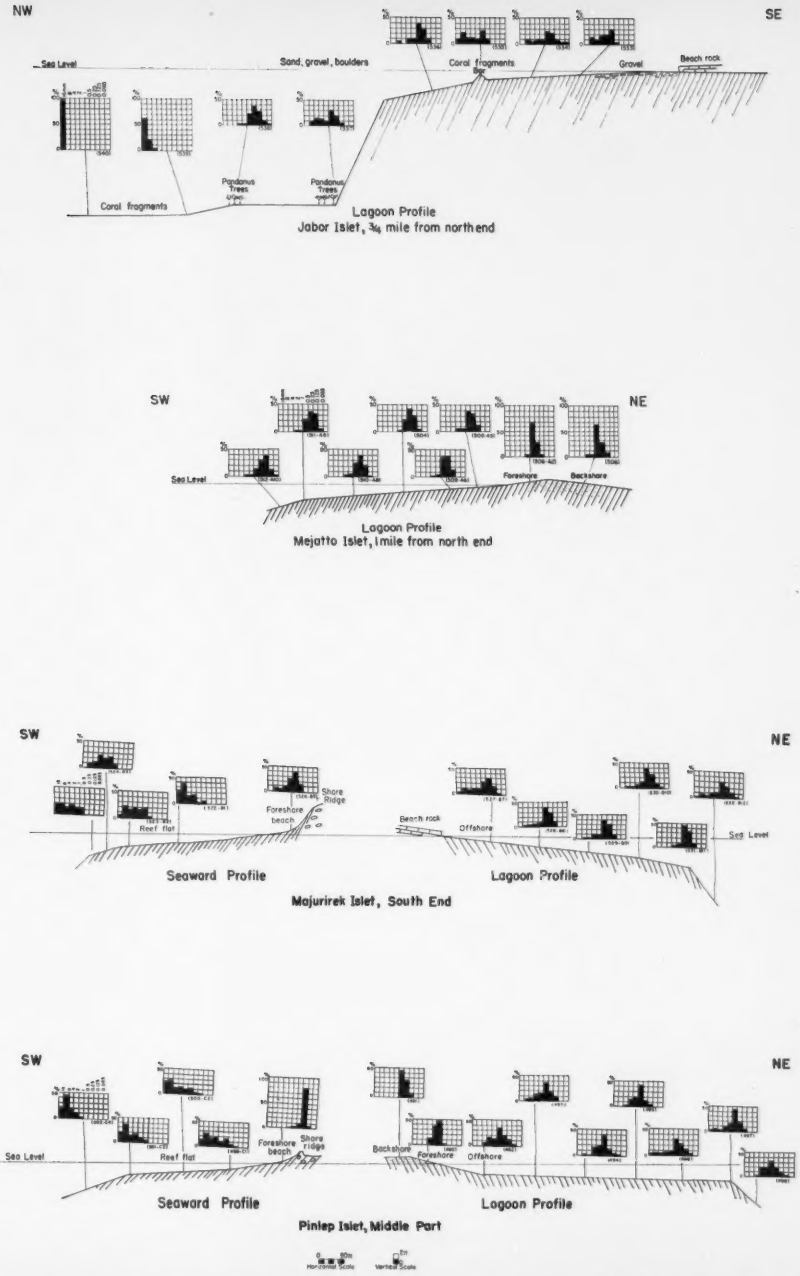

FIG. 4.—Profiles of reef and lagoon floor bordering islets with accompanying sediment analyses.

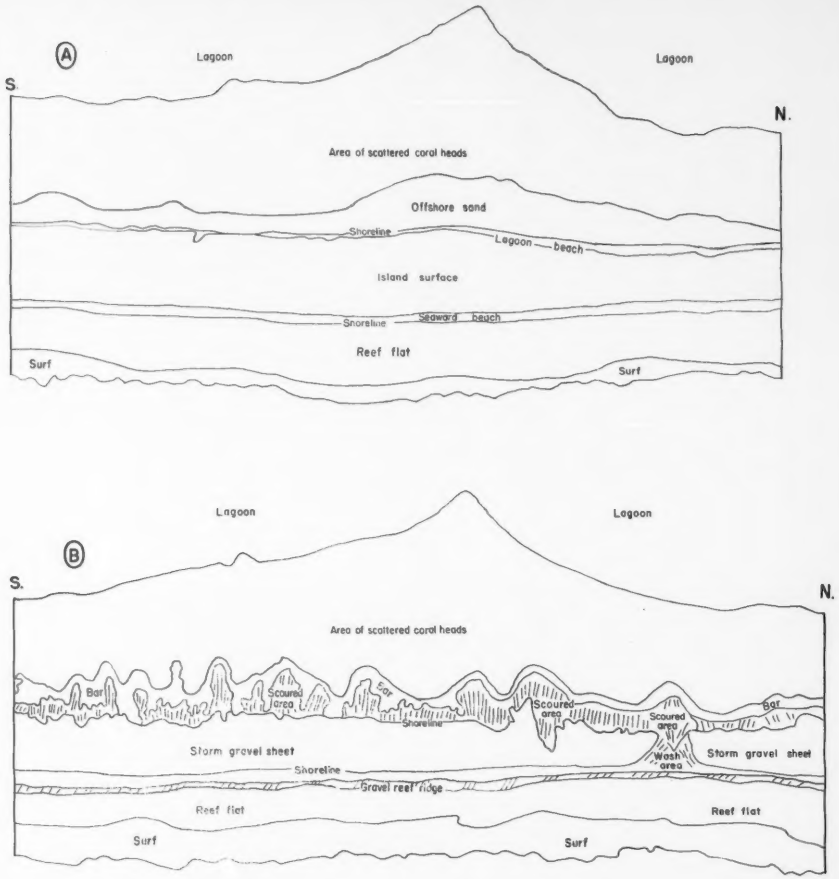

FIG. 5.—Maps showing southern part (about ¾-mile long) of Mejatto Islet in 1944 (A) and in 1958 (B). Traced from aerial photographs of the Office of Naval Research. Map compiled in 1958 shows addition of gravel ridge on reef flat, gravel sheet on islet, areas of scour and of looping bars offshore—all results of typhoon.

at Kapingamarangi Atoll (McKee, Chronic, and Leopold, 1959), but which were deposited under normal conditions of waves and currents, suggests that the proportionately smaller amount of foraminifera in the very near shore waters and their correspondingly greater numbers far out from shore at Jaluit are direct results of redistribution by the typhoon. The relatively larger amount of coral debris near shore off Jaluit may also be a result of the typhoon.

Studies of bottom sediments on seaward sides of islets on Jaluit Atoll were attempted for comparative purposes. Sand was absent on the seaward sides of Jabor and Mejatto Islets, probably having been removed by the storm waters that swept from the reef flats across the islets. On Pinlep and Majurirek Islets, where the storm was less intense, fine sediment of the reef flat was

poorly sorted and relatively coarse, and median diameters were larger than sand size (fig. 4). The sediment was composed largely of coral fragments, and contained some broken mollusk shells and *Halimeda* segments, but no foraminifers. Essentially no fine sand was present.

Sand beaches do not now exist along much of the lagoonal sides of Jabor and Mejatto Islets. Aerial photographs taken since typhoon Ophelia show that much of the area formerly occupied by beaches on these islets is now scoured to reef rock surface. Sand deposits now form loops or bars in the offshore waters, each bar appearing as a half-circle, convex outward. This pattern forms a conspicuous feature along the southern part of Mejatto Islet (fig. 5). Gentle lagoonward slopes and steeper shoreward slopes on these bars, as seen in the photographs, are believed to have resulted from the gradual reworking of the sand masses by incoming waves off the lagoon.

EFFECTS OF TYPHOON ON GROUND WATER

Although no studies of ground water had been made on Jaluit Atoll prior to typhoon Ophelia, fresh-water lenses are known to have existed on all but the very small islets, as shown by dug wells which for many years furnished potable water. The quality of this water probably was similar, in general, to that reported from fresh-water lenses on other Pacific limestone islands. Storm waves of the typhoon, however, submerged many of the islets, so their supply of ground water was locally contaminated and its composition greatly altered by mixing with sea water.

Samples of ground water were collected for analysis in May 1958, about four months after the storm. These samples were obtained from four wells in use prior to the typhoon, three test wells dug at the time of sampling, one pool in a scour trench cut by the typhoon, and one mangrove pond. The three test wells and one of the original wells were on Jabor and Mejatto Islets, each of which was covered by a sheet of water during the typhoon; the other three old wells were on islets not inundated (Pinlep, Majurirek, Kinajon) and their waters apparently were little changed as a result of the storm.

Temperature and pH readings were taken at each of the wells examined (table 1). Temperatures ranged from 82°F to 87°F. These were read near the middle of the day and probably were relatively high because all the wells were in the open, unprotected from the sun's heat. Well waters ranged from about 6.0 to 7.0 in pH, the higher figures representing wells on islets known to have been inundated by typhoon-driven sea water.

TABLE 1.—*Readings of pH and temperature of well water*

Well description	pH	Temperature (°F)
Majurirek	6.0	86
Pinlep	6.5	83
Mejatto well No. 1	7.0	85
Mejatto well No. 2	7.0	87
Jabor well No. 2	7.0	82

Hardness of water (table 2) is less than 500 ppm in each of the three samples from islets not inundated by storm waves; in all others it is greater than 1200 ppm, apparently because of contamination by sea water. Thus, the low calcium and low magnesium content in well waters of Pinlep, Majurirek, and Kinajon Islets probably is normal for fresh water on these islets; water of similar composition is reported from fresh-water lenses in the northern Marshall Islands (Arnow, 1954, p. 7) and on Kapingamarangi Atoll (McKee, 1958, p. 267). The amount of calcium and magnesium in such water is higher than is average for fresh water and is probably due to the solution of limestone and lime sand of the islets.

Well water from Jabor and Mejatto Islets (table 2), which were covered by sea water during the typhoon, is much higher in percentage concentration in both calcium (more than 200 ppm) and magnesium (more than 170 ppm) than water from the islets cited above that were not inundated. Furthermore, the amount of magnesium in all but one of the Jabor and Mejatto samples is equal to, or greater than the amount of calcium. This is probably also a result of contamination, for normal sea water has proportionately more magnesium than calcium, whereas most fresh water has the reverse.

The extent of contamination by sea water is illustrated by analyses of sulfate (SO_4) and

TABLE 2.—*Analyses of water samples from original dug wells and test wells and comparative analyses for a mangrove pond and for normal sea water. (Analyses by U. S. Geological Survey)*

Location of well	Parts per million								
	Ca	Mg	Na	K	SO_4	Cl	Hardness as $CaCO_3$	Percent sodium	Specific gravity
Majurirek, 190 ft from lagoon beach crest W. end of islet center	102	17	138	11	45	238	324	47	1.000
Pinlep, well, 190 ft from lagoon beach crest	53	8.8	62	5.8	26	62	168	43	1.000
Kinajon, 130 ft from low water, 99 ft from ridge on southwest islet	142	27	316	27	102	570	466	58	1.000
Mejatto, well No. 1, 174 ft from seaward gravel ridge	285	529	4,900	214	1,370	8,700	2,890	73	1.011
Mejatto, well No. 2, 301 ft from lagoon beach crest	362	888	7,550	252	1,910	13,000	4,550	77	1.017
Jabor, well No. 1, 130 ft from lagoon beach crest	223	242	1,990	84	535	3,390	1,550	72	1.004
Jabor, well No. 2, 265 ft from lagoon beach crest	207	174	1,500	84	473	2,890	1,230	71	1.003
Mejatto, water from scour pool on lagoon side of islet behind beach crest	159	415	3,570	147	1,010	5,950	2,100	77	1.008
Majurirek, water from mangrove pond	376	1,090	8,700	376	2,300	14,900	5,420	76	1.020
Normal sea water	400	1,270	10,560	380	2,650	18,980	6,215	79	

chloride (Cl) for well water from Jabor and Mejatto Islets (table 2). The water on Mejatto has a combined sulfate and chloride content ranging from 10,000 to 15,000 ppm; that on Jabor ranges from 4000 to 5000 ppm. The U. S. Public Health Service recommends 250 ppm of sulfate and 250 ppm of chloride as the upper concentration limits for water used in normal domestic consumption.

A summary of available data on the ground water resources of Jaluit Atoll after typhoon Ophelia is as follows. The fresh-water lenses on those islets not covered by storm waters seem to be normal for Pacific atolls and probably were little or not affected by the typhoon. The fresh-water lenses on Jabor and Mejatto show a sulfate-chloride content far too high for drinking purposes, and will require a considerable period of dilution by rainwater to again become potable. How long a period of "freshening" will be needed with the present annual precipitation of about 200 inches is not known, but samples from the wells in question, if collected and analysed periodically, should give significant information.

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FIGS. 1–6, SEPTEMBER, 1959

DIAGENESIS IN MISSISSIPPIAN CALCILUTITES AND PSEUDOBRECCIAS[1]

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ABSTRACT

The diagenetic histories of some Mississippian calcilutites from England and Wales are considered in the light of fabric criteria recently put forward by the author (1958). It is concluded that calcite-mudstone was, at the time of its cementation, a mud of discrete calcite crystals, only slightly finer-grained than the present mosaic. Whether the crystals had earlier inverted from aragonite it is impossible at present to say. Calcite-siltstones seem to have had a similar history, and evidence for cementation of original calcite crystals is to be seen in some examples. Criteria are presented for the distinction of mudstone aggregates (Bahamite "grains," faecal pellets, etc.) from relics of calcite-mudstone remaining after grain growth. Distinction is based on a study in thin section of the shapes of the finer-grained areas and their distribution, and of the fabrics of the intervening coarser mosaics. The fabrics of pseudobreccia are described in detail and are shown to be the result of patchily developed grain growth.

INTRODUCTION

A fuller understanding of the calcilutites has long been hindered by the occurrence in these fine-grained limestones of coarse mosaics which are not rim-cemented detritus (Appendix). Some of these coarse mosaics are crystalloblastic replacements of the original, cemented sediment, some are granular cement (Appendix), and others are drusy infillings of cavities. It is the purpose of this paper to present criteria by which the various kinds of mosaic can be recognised and to apply these criteria to the examination of some Mississippian calcilutites (grain diameters less than 60μ).

In any limestone the *matrix* is an accumulation of one or more types of grain mosaic. In the Mississippian (Dinantian) limestones of England and Wales, three main types of mosaic can be distinguished and these make up all the limestone matrix, except for small amounts of secondary mosaics which have grown locally as a result of strain. These latter mosaics are quantitatively unimportant and are not considered here. The three dominant mosaics are (1) granular cement and drusy mosaic, (2) rim-cemented single crystals, and (3) grain growth mosaic. They have been dealt with at length elsewhere (Bathurst, 1958) and only brief notes are included here in the Appendix.

[1] Manuscript received February 2, 1959.

The calcilutites in their simplest form are rim-cemented carbonate muds or silts. Commonly, however, the original sediment was composed of aggregates of mud or silt, either faecal pellets or "grains" similar to Illing's (1954) Bahaman sands. It is here that confusion has arisen in the past, because in thin section the aggregates, lying in a coarse granular cement, can be mistaken for relics of calcilutite remaining after grain growth. Calcilutite may also be complicated by the occurrence within it of rounded patches of coarser mosaic. These, also, are commonly mistaken for crystalloblastic mosaic; but detailed examination often reveals that they are drusy infillings of cavities. Yet, in some of the limestones, grain growth mosaic is common and in one limestone it is an essential component. This is the pseudobreccia where the "fragments" are masses of grain growth mosaic which lie in a "matrix" of less altered limestone.

FABRIC EVOLUTION OF CALCITE-MUDSTONE

Fabric

In thin section calcite-mudstone appears as a non-porous mosaic of roughly equant grains.[2] It is the finest of all the non-skeletal mosaics. Grain diameters lie almost entirely within the range 0.5–4μ.

[2] "Grain" as used in this paper refers to single crystals.

Deposition

Much of the original sediment was mechanically deposited. Sander's (1936, p. 31; 1951, p. 1) definition of mechanical deposition is followed. He applies it to "particles that have been developed in some way, entirely before the depositing process, for example, as detritus, crystals, oölite grains, etc." Diagnostic criteria are: (i) preferred orientation of skeletal particles, pellets, etc., with longest or platy axes sub-parallel not only to each other but also to any lamination; (ii) laminae differing in content of skeletal particles, pellets, quartz grains, etc.; (iii) undoubted detrital[3] particles (*e.g.* foraminifera, quartz grains) surrounded in three dimensions by calcite-mudstone; (iv) uniformity of texture—unlike grain growth mosaic or chemically deposited mosaic (Appendix); and (v) where calcite-mudstone occupies the lower part of a cavity it has a smooth, more or less flat, upper surface. Only in exceptional cases is it obvious that the mudstone is not wholly a product of mechanical deposition, for example in algal growth-forms. Any apparently homogeneous mudstone may have been initially an accumulation of faecal pellets which lost their individualites on compaction like those off the Florida coast (Ginsburg, 1957, p. 81).

Aragonite- or Calcite-mud

If the original mud was aragonite, then all trace of the needle-shaped grains has been lost. Inversion to calcite would have caused a volume increase of more than 8 percent. If this happened, then it must have been before the grains were rim-cemented into a rigid fabric because there are no signs of such a drastic change. When aragonite ooliths in Great Salt Lake, Utah, invert to calcite they develop superficial bumps (Eardley, 1938, p. 1384). Cullis (1904, p. 400–403) described replacement of a rigid, drusy, aragonite fabric by calcite in the deeper parts of the Funafuti reef, but this involved an increase in grain diameter of from 10 to 100 times. A similar increase accompanied calcitisation of aragonite ooliths in the Pleistocene Miami Oolite (Ginsburg, 1957, p. 96–97).

Wood (1941, p. 194) concluded that the original mud was probably calcite on the grounds that tubes which had been occupied by algal filaments now have sharp walls in contact with the mudstone. He believed that a change in the surrounding mud from aragonite to calcite would have blurred this sharpness. Though inversion in a compact biogenic deposit around empty filaments might well blur their boundaries, such damage may have been resisted here by the drusy calcite which fills the tubes.

Although there is no fabric evidence for an aragonite origin, it would be surprising, judging by the importance of this mineral in modern carbonate muds, if bacterial, meteorological, or other influences had not brought about some aragonite deposition in the Mississippian seas.

Elimination of Pores

The mud was presumably changed to a non-porous mosaic by solution of supersoluble particles and redeposition, and by solution transfer (Appendix). Cementation by carbonate brought from outside the mud is unlikely in view of the low permeability of mud. Plastic deformation under load needs more energy than solution transfer and the grains show no twinning. Viscous flow and solid diffusion are also processes using more energy than would be needed for pore elimination in a mixture of grains and solution.

The Original Grain Size

It seems clear that the original mud grains were no larger than those found now in the mosaic. There is no reason to suspect derivation by fracture from larger grains. The various kinds of secondary origin involving an increase in grain size are also out of the question, because the mudstone grains are nearly as small as detrital carbonate grains can be. Neither primary recrystallisation (the seeding of new grains in a strained lattice) nor grain growth would have yielded such a uniform product—uniform throughout Mississippian time and in geographical distribution. Solution transfer (Appendix), involving only solution and redeposition, could have done little but modify grain shape; and the solution and redeposition of small supersoluble grains could have had only a very slight effect.

In discussing the precipitation of carbonate muds in the form of algal dust, Wood

[3] "Detrital" and "detritus" are used here as adjective and noun for mechanically deposited sediment.

(1941, p. 194) thought it likely that the present grain sizes scarcely differ from the original, because of the similarity of grain sizes in modern cultured algal dust and also of assumed algal dust inside Mississippian ooliths.

The calcite-mudstones are also very similar in grain size to aragonite mud, from which some of them may have evolved. Vaughan (1917) found that the aragonite needles of the Bahaman and Florida sea floor deposits have shortest and longest diameters of 0.5μ and 3μ. Illing (1954, p. 27) gives the fundamental particle sizes on the Bahama Banks as 1μ–4μ. An electron photomicrograph of Bahaman aragonite ooze (Newell and Rigby, 1957, Pl. XVI, fig. 2) shows needles with dimensions from 0.1:0.4μ to at least 0.4:2.5μ.

The range of grain diameters in the finer (coccolith) fraction in the Chalk of northern Europe is 0.5μ–4μ (M. Black, 1953, p. lxxxi), the same as in the Mississippian mudstone.

It seems possible, therefore, that grains of 0.5μ are about the smallest particles of calcium carbonate that are stable in a buried sediment, but that the original grain size of the mudstone, while having much the same upper limit as it has now, may have had a lower limit of, perhaps, a fifth of the present value. Solution and redeposition of supersoluble grains, and solution transfer, could have removed the smallest grains after burial.

Stability of the Mudstone Fabric

The uniformity of grain size throughout the Mississippian calcite-mudstones points to the existence of a universal threshold state at which fabric evolution stops and beyond which it can, but need not, continue (Bathurst, 1958, p. 30). The factor which has determined this grain size must be the fabric of the original mud. Cementation gave this a stability which has enabled many calcite-mudstones to remain unaltered for tens of millions of years. The threshold state marks the change from the action of solution-deposition to that of grain growth.

FABRIC EVOLUTION OF CALCITE-SILTSTONE

Little is known about these sediments in the Mississippian limestones. Banerjee (1959) has recorded upper Viséan calcite-siltstones from North Wales. Eleven thin sections of these show a grain size range from about 5–35μ, differing only slightly from one sample to another. Fossiliferous calcite-siltstones (previously described as calcite-mudstones) form the more or less lenticular knoll reefs of the English Midlands. They have a range of 2.5–14μ. Identical material has been seen in samples from Feltrim knoll reef near Dublin, Eire. Pray (1958) has described Mississippian knoll reefs of similar type from New Mexico and he gives the grain size of the sediment as about 10μ. Knoll reefs are commonly about 300 feet thick, with original slopes of 40°–50°, and the accumulation of such large quantites of carbonate silt, so different from the normal calcite-mudstone, is a matter of great interest. Unfortunately, definite information on its genesis is lacking.

The same criteria for mechnical deposition are visible in the siltstones as in the mudstones. Algal origin is less likely, because recognisable algal deposits have the grain size range of calcite mudstone. Definitive evidence for the original mineralogical state has not yet been produced. It is, however, possible to resolve the intergranular boundaries in the coarser siltstones in sufficient detail to show that they are composed of a number of plane surfaces. This type of boundary is an important criterion for the recognition of a rim-cemented mosaic. The conclusion follows that, whatever the initial mineralogy of the sediment, the most recent diagenetic act recorded by the fabric was rim-cementation of discrete calcite crystals.

There is no reason to suppose that the original sediment was anything but a slightly finer silt. The same arguments apply here as for the mudstone. Calcite-siltstones appear to have been quite unaffected by grain growth.

CALCITE-MUD AGGREGATES AND DRUSY CAVITIES: DISTINCTION FROM GRAIN GROWTH FABRICS

Part of the problem here (discussed by Beales, 1956, p. 864–866) is the distinction between mechanically deposited mud aggregates and the more or less rounded,clot-like, relics of the recrystallisation of calcite-mudstone. Though in many limestones this distinction is straightforward there are yet a number of examples where the original na-

ture of the material is far from obvious and about which interpretation has been controversial. Confusion arises largely because the mosaics formed by chemical deposition and by grain growth resemble each other superficially—and this apparent similarity is nowhere more disconcerting than in the "clotted" calcilutites. These fine-grained limestones include such varieties as the clotted limestones of Wood (1941), the *structure grumuleuse* and *pseudoolithe* limestone of French and Belgain writers, the German *Krümelgefüge*, the pellet limestones of Cayeux (1935, fig. 60) and M. Black (1938, p. 169), the Bahamites of Beales (1958), and the enigmatic pseudobreccias.

The basic pattern in all these limestones is a twofold division into masses of calcite-mudstone or calcite-siltstone wholly or partly embedded in a coarser calcite mosaic. Many of the patterns are not obviously detrital in origin and a good deal of uncertainty still exists regarding their causes and, particularly, the amount of recrystallisation which has taken place *in situ*. *Structure grumuleuse* is a case in point and Cayeux's description of it (1935, p. 271) is still the most complete while being happily free from genetic implication. He wrote, ". . . elle montre de tout petits éléments calcaires, á pâte extrêmement fine, se détachant en gris sombre, de forme générale globuleuse ou irrégulière, dont les contours ne sont jamais franchement arrêtés, et sans différenciation d'aucune sorte. Ces matériaux, dont la microstructure est invariablement cryptocristalline, sont plongés dans une gangue de calcite incolore et grenue." Cayeux recognized that the masses of calcite-mudstone may be formed in more than one way, but his Plate 18 (*op. cit.*) shows clearly the difficulty of separating the detrital from the diagenetic.

Another problem with regard to recognition arises from the common occurrence in calcite-mudstones and calcite-siltstones of scattered rounded islands of coarse mosaic. These are generally attributed to patchy recrystallisation of the mudstone though they are usually drusy cavities.

A mud aggregate is defined here an any aggregate of mud grains, usually having the size of a sand or silt particle, which has been mechanically deposited. Initially the aggregate may have been a faecal pellet (Eardley, 1938, p. 1401; Illing, 1954, p. 24), or a rounded, sub-spherical aggregate of mud grains cemented originally by aragonite with no signs of organic control (as Illing's Bahaman sands, 1954, p. 27 *et seq.* which lithify to yield the Bahamites of Beales, 1958), or a fragment of algal precipitate (Wood, 1941, p. 192; George, 1954, p. 306 and 1956, p. 311; Lowenstam, 1955; Lowenstam and Epstein, 1957), or a spherical or ovoid growth form of a calcareous alga (Anderson, 1950, p. 8, 14). Local occurrences of mudstone fragments which are abraded remains of older limestones have been described by Kaisin (1926, p. 1271) and Cayeux (1935, Pl. VII), but these clastic aggregates were not seen in the limestones, except locally as resediment in some cavities in knoll reefs.

Considering first of all the aggregates, criteria favouring mechanical deposition present no difficulty. They are: a preferred orientation of longer axes parallel to the bedding; layers differing in aggregate size; admixture of aggregates with skeletal particles or ooliths; contact packing of the aggregates; homogeneity of aggregate size.

Criteria consistent with relic structure (figs. 1 and 2) are: irregular shapes of clots; much variation in clot size; clots floating out of contact in a coarser mosaic which does not appear to be detrital; merging of aggregates (Cayeux, 1935, figs. 67, 68; Beales, 1956, p. 864).

The shape of an aggregate may be a useful indication of its origin. Some faecal pellets and algal growth-forms are typically elongate and rounded; some algal forms are concentrically layered; some clastic particles are angular. It is not safe, however, to assume that a smooth, unembayed, surface was once formed by the processes of abrasion and accretion; grain growth can advance on a remarkably smooth front in mudstone, though usually some sawtoothed contacts are visible.

The coarse matrix between the aggregates and clots occurs in three forms, and it is always coarser than the mosaic within the aggregates. It may be the product either of rim-cementation of a coarse mud or silt, or of granular cementation, or of grain growth.

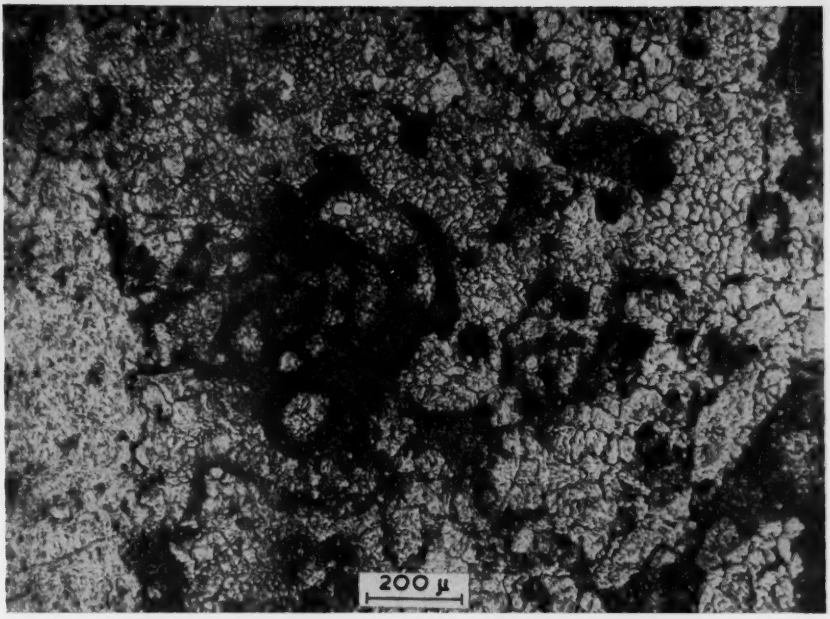

FIG. 1.—Relics of fossiliferous calcite-mudstone in grain growth mosaic.
D_1 zone, Graig Quarry, Denbigh, N. Wales.

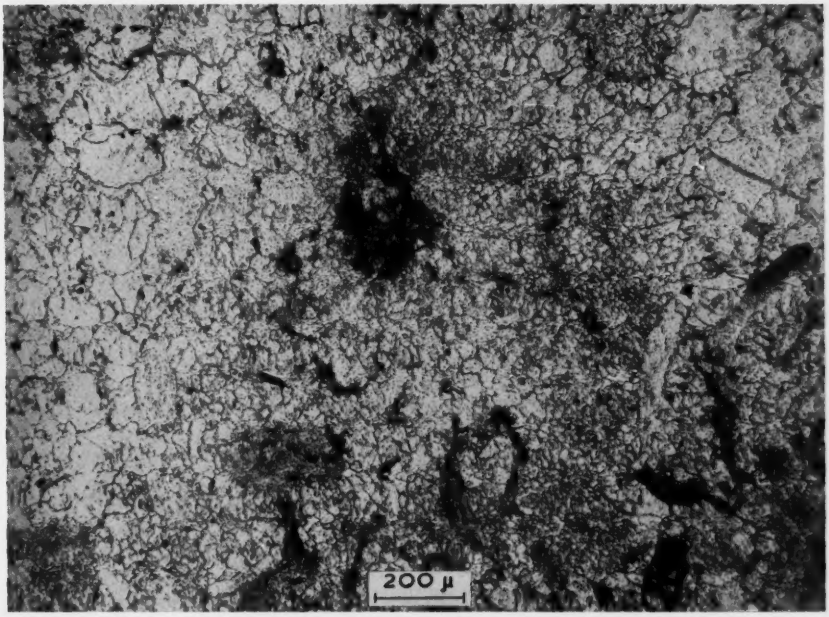

FIG. 2.—Relics of fossiliferous calcite-mudstone in grain growth mosaic.
D_1 zone, Meal Bank Quarry, Ingleton, Yorks.

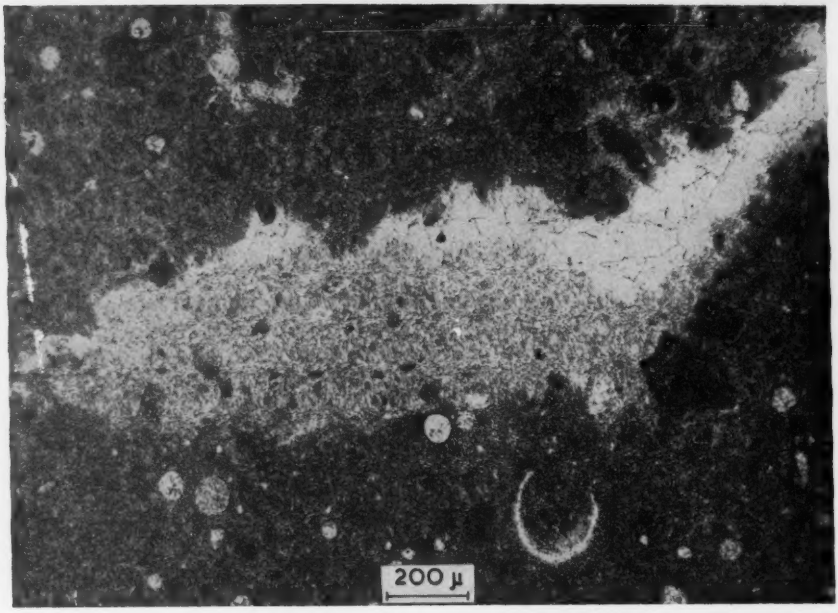

FIG. 3.—Cavity in calcite-mudstone filled with an internal sediment of calcite-siltstone overlain by drusy mosaic. D_1 zone, Tan Dinas Quarry, Careg-onen, near Penmon, Anglesey.

Rim-cementation of a mud or silt is indicated by the homogeneity of the grain size and plane integranular boundaries. These sediments are also spread over the floors of some cavities.

In a matrix of granular cement, plane intergranular boundaries are a useful criterion with, perhaps, an increase in grain size away from the wall and preferred orientations of optic axes and longest axes normal to the wall.

A matrix of grain growth mosaic can be recognised by the absence of the foregoing characteristics and by curved to consertal intergranular boundaries. Grain shape may be very irregular and the grain size is patchily and irregularly distributed. In some limestones there is a transition from coarse matrix, through intermingling matrix and mudstone, to mudstone so that no simple boundary can be seen. On the other hand, the presence of a smooth boundary between coarse mosaic and aggregate can be a result either of cementation or of grain growth. Boundaries of pellets and foraminifera, for example, show a marked resistance to grain growth even where it has selectively altered the mudstone matrix.

Turning now to the islands of coarse mosaic in a matrix of calcite-mudstone or calcite-siltstone (figs. 3 and 4), these have a great variety of shape and have clearly been formed in a number of ways. Nevertheless, by applying to them the criteria for chemical deposition and grain growth, it can be shown that in nearly all cases the islands are filled with drusy mosaic. Indeed, grain growth rarely gives rise to sharply defined islands of coarse mosaic; even in pseudobreccias these are more apparent than real.

The processes causing the initial cavities probably include solution of aragonite debris, decompostion of organic tissue, burrowing, the evolution of gas in soft mud, incomplete sedimentation in voids between sand grains, and internal erosion (Sander,

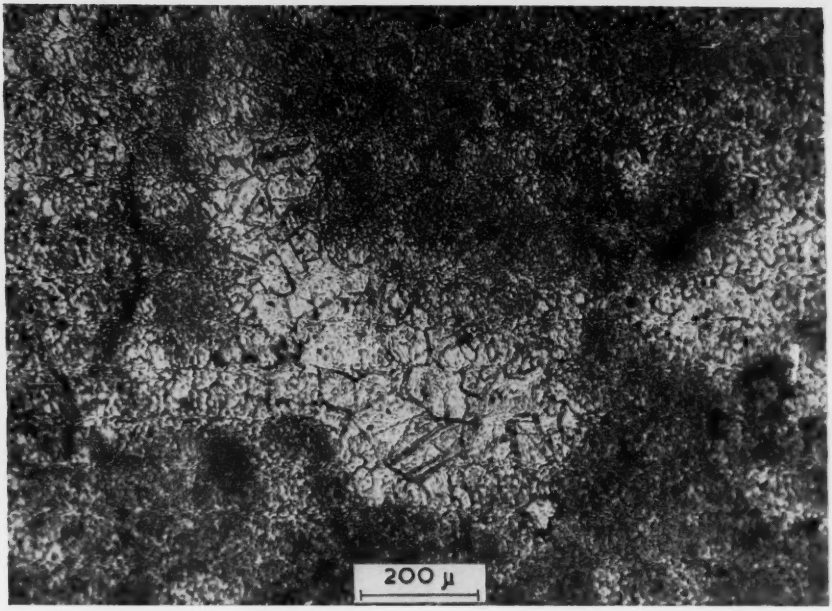

FIG. 4.—Cavity in calcite-mudstone filled with drusy calcite. Locality unknown.

1936, p. 50, 94; 1951, p. 26, 58) by migrating pore water. In a few instances evidence for the earlier existence of a cavity is also shown where the bottoms of the cavities are covered by a deposit of siltstone or other sediment (fig. 3). Recrystallisation of calcite-mudstone (grain growth) seems to have been much less common than is usually supposed.

PSEUDOBRECCIAS

The microfabrics of these spotted limestones are excellent illustrations of patchy alteration by grain growth, and the pseudobreccias are also interesting because they show a wide range of fabrics caused by this process. The descriptions are based on samples from Ingleton, in Yorkshire, and from North Wales.

The Mississippian pseudobreccias were fully described by Dixon and Vaughan (1911, p. 507–511). They distinguished two components, both of which are visible to the naked eye. Their "fragments" are irregularly shaped lumps of coarse calcite mosaic, usually between 1 mm and 20 mm in diameter, and are dark grey in the hand specimen. They lie in the finer, pale-grey, "groundmass" of calcite-mudstone. In thin section (fig. 5) the "fragments" appear light and the "groundmass" dark. The outlines of the "fragments" are clearly not typical of detrital particles (Dixon and Vaughan, 1911, p. 507–508). In the field, however, they may differ markedly in size from one bed to another and in some beds they are graded, being smaller towards the top. It is possible that an original, detrital fabric may have influenced the pattern of subsequent recrystallisation.

In discussing the genesis of pseudobreccia, Dixon and Vaughan suggested that some of the skeletal debris in the original, unconsolidated sediment had dissolved in the pore water; and that this solute had been reprecipitated in the pores of certain limited regions which now appear as the coarser grained patches or "fragments." As a result of this process, the finer "groundmass" was, they supposed, left relatively unfossilifer-

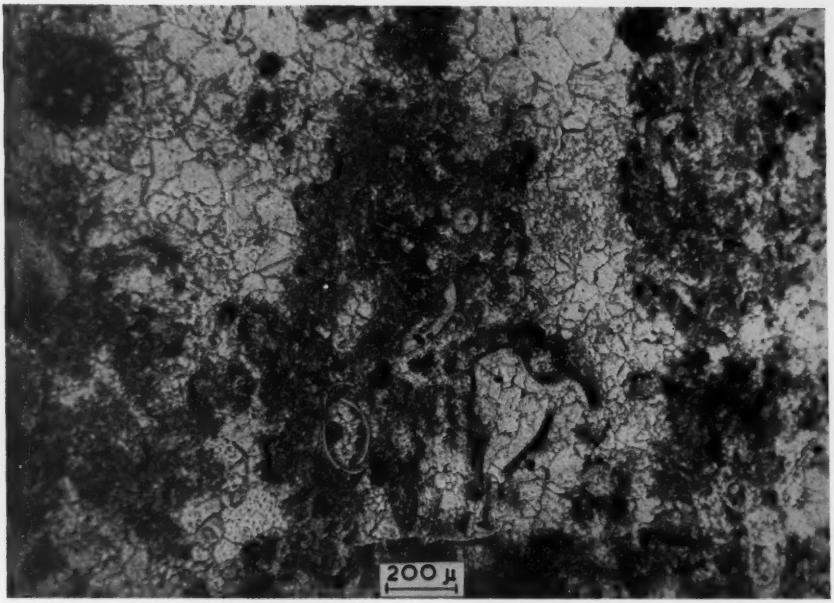

FIG. 5.—Pseudobreccia: patches of unaltered fossiliferous calcite-mudstone interspersed with grain growth mosaic. D_1 zone, Meal Bank Quarry, Ingleton, Yorks.

ous, the few fossils which remained being vague in outline owing to their partial solution; the "fragments," on the other hand, still have their original skeletal content.

However, detailed fabric analysis, based on the criteria proposed by Bathurst (1958), does not support this hypothesis, but indicates that the "fragments" are islands of more advanced grain growth mosaic embedded in a "groundmass" of less altered limestone.

First of all, the "groundmass" is seen to have, instead, a higher skeletal content than the "fragments." A further objection is the apparent impossibility of changing a carbonate mud into a much coarser fabric, with grain diameters enlarged by a factor of about 20, merely by chemical redeposition in the pores of the same mud.

Examination, in thin section, of the less altered parts of pseudobreccias shows that two types of carbonate sediment have been recrystallised—a skeleton-rich calcite-mudstone (fig. 2) and a fine carbonate sand (fig. 6). Each contains foraminifera, crinoid fragments, brachiopod fragments, *Koninckopora*, etc. In the sand there are Bahamite "grains" (Beales, 1958), and on some of the sand particles there are traces of first generation cement. These small cement grains have plane intergranular boundaries and, commonly, crystal faces overgrown and preserved by later, bigger grains.

Only the fine mosaic (0.5–4μ) has been replaced. This includes calcite-mudstone, the walls of foraminifera, *Konickopora*, and Bahamite particles. The contact between the fine mosaic and the coarse, secondary mosaic may be gradational and vague, or it may be perfectly sharp. The crinoid fragments have not been replaced. On the contrary, they are usually enlarged and show clear secondary rims about a dusty host. Patches of wholly unaltered sediment are rare; the "groundmass" generally shows some degree of coarsening.

Stages of replacement in the skeleton-rich calcite-mudstone are: (i) patchy coarsening

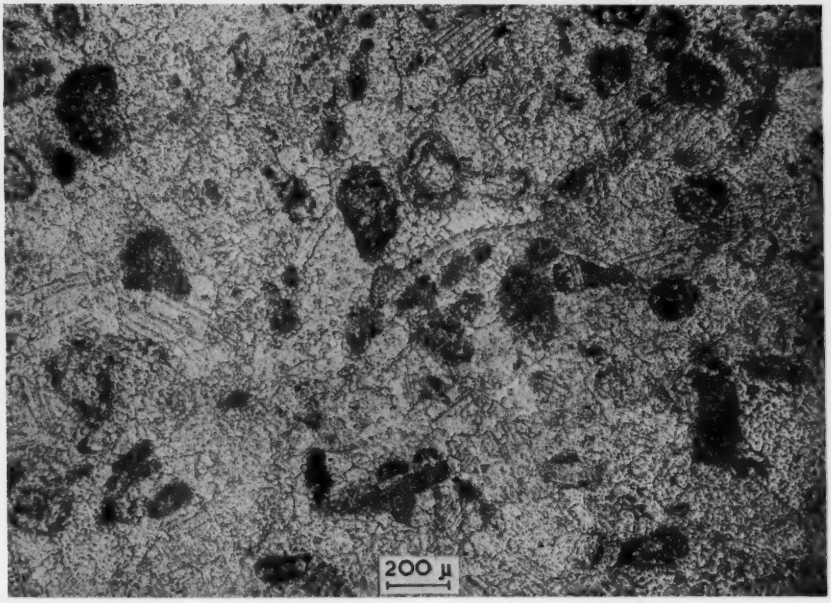

FIG. 6.—Pseudobreccia: residual sand particles in a grain growth mosaic which contains scattered relics of unaltered cement. Di zone, Meal Bank Quarry, Ingleton, Yorks.

of the mudstone with embayment of the remaining relics of mudstone by plane-sided marginal grains of the coarse mosaic giving a more or less sawtoothed contact: the skeletal and Bahamite particles are left with their surfaces entire and with no internal alteration; (ii) sawtoothed embayment of the skeletal and Bahamite particles by coarse mosaic; (iii) general patchy development of coarse grains within the skeletal and Bahamite particles; (iv) total replacement of all remaining fine mosaic.

Many of the grains in the coarse mosaic have diameters (in section) between 50–100μ. Those smaller than about 7μ are indistinguishable from the original fine mosaic.

The alteration of the calcite-sand followed the same course, beginning at (ii).

Floating detrital particles and patches of mudstone are commonly so completely isolated from each other that it must be assumed that they are really out of contact in three dimensions. As the coarse mosaic is clearly not detrital (i.e. rim-cemented single crystals), some earlier detrital fabric must have been replaced.

The plane-sided, or sawtoothed, embayments appear to be incompletely developed rhombohedra. Occasionally whole rhombs of calcite are visible, embedded in calcite-mudstone or in a Bahamite particle. This phenomenon occurs not only in pseudobreccia but in Mississippian limestones generally, where they have been recrystallised by grain growth. It would seem to mark an approach to a crystal form having maximum stability with regard to the enveloping fine-grained mosaic. The sawtooth contact in no way resembles a stylolite. As in the case of crinoid fragments which have been enlarged by grain growth at the expense of calcite-mudstone (Bathurst, 1958, p. 29), this is clearly a process of addition to pre-existing, coarse grains and not one of subtraction.

All the grain-growth mosaics show the characteristic fabrics: irregularity of grain shape and of intergranular boundaries, heterogeneity and patchiness of grain size

(Appendix). Patches of uniform sugary mosaic are a common early stage, having grain diameters more or less in the range 50–150μ.

The fate of the cement in the calcite-sand is of interest. So far, in the study of grain growth in various Mississippian limestones, no evidence has been found for the replacement of cement, except in the rim-cemented mudstone mosaic (Bathurst, 1958, p. 31). Pure cement appears to be stable. In these altered calcite-sandstones, scattered plane intergranular boundaries are visible in the coarse mosaic and these are presumably relics of the cement fabric. The preservation of the crinoids, must, it seems, be placed in the same category. Recrystallisation of cement mosaic is certainly not inhibited by grain size, because many grain growth mosaics are coarser. On the other hand, both crinoid skeletons and cement were deposited under conditions of low lattice strain and were, presumably, always more stable than the calcite-mudstone. Though this is rim-cemented, it has a high ratio of surface to volume and, consequently, a higher surface energy than the much coarser cement mosaic. The continuation of grain growth to a stage where mosaics coarser than many cements are evolved might be a result of delayed release of elastic strain.

The reason for the patchy development of grain growth in pseudobreccia is obscure. Grain growth is known to be inhibited by impurities in the lattice and by the presence of voids. It may be that the patchy recrystallisation represents an original heterogeneity in one or both of these factors, but this problem remains to be solved.

Many pseudobreccias have been dolomitised (*e.g.* in South Wales) and an interesting conclusion can be drawn about the time of metasomatism. Since dolomitisation played no part in the formation of the pseudobreccia fabric, its preferential replacement of the "groundmass" means that that this must have been differentiated before dolomitisation occurred. In other words, dolomitisation took place not only after the calcium carbonate mud had become a mudstone, but after it had proceeded to the further stage (possibly millions of years later) of grain growth.

APPENDIX

Some notes on the various processes referred to in the text are given below. A full treatment of them, their mechanisms and resultant fabrics, has been published elsewhere (Bathurst, 1958).

Chemical Deposition

This refers to the growth of crystals, from solution, on a free surface, as distinct from precipitation of crystals as a suspension.

"Drusy mosaic" is a convenient name for grain mosaics which have been deposited on the walls of cavities. "Cement" is more commonly used when deposition has taken place in the pores of a detrital sediment. "Drusy mosaic" and "granular cement" have been used by Bathurst (*op. cit.*, p. 14) to describe mosaics which have grown on multigranular surfaces containing many available nuclei (*e.g.*, the wall of a foraminifer). These mosaics are distinguished from "rim-cement" (*op. cit.*, p. 21) which grows as a syntaxial extension of a detrital single crystal (*e.g.* a crinoid fragment). Whereas drusy mosaic and granular cement are entirely chemically deposited, the mosaic formed by rim-cementation consists of grains each of which has a core composed of the detrital host.

The growth of chemically deposited mosaics is partly controlled by competition between adjacent grains. Those grains with their axes of fastest growth oriented normally to the wall overgrow those not so oriented. Consequently, the number of surviving grains decreases away from the wall, and the grain size of the mosaic increases away from the wall. At the same time, preferred orientations develop with optic axes and longest axes normal to the wall. A further development is that of plane, intergranular boundaries in the mosaic. This is a normal occurrence among competing crystals, whether they have grown by chemical deposition or from a melt. Adjacent faces join along a line and, where their growth rates are related by a constant, this line moves in a plane.

Solution Transfer

This is a translation of the German *Lösungsumsatz*. It refers to the solution of

detrital particles around their points of contact where elastic strain and solubility are enhanced (pressure solution), followed by redeposition on less strained particle surfaces.

Grain Growth

This is the name given originally by metallurgists to a process of grain enlargement which acts in the solid state. Where the porosity has been sufficiently reduced (in limestones by cementation or solution transfer) to give adequate intergranular boundaries, these migrate so that, in general, large grains replace smaller ones and the mosaic becomes gradually coarser. The driving force is believed to come partly from differences of surface tension across the intergranular boundary and partly from variations of elastic strain between neighbouring grains. Strain of this kind is never wholly absent from a lattice and is always associated with intergranular boundaries where the lattices are distorted to adjust for misorientation across the boundary.

Grain growth mosaics have fabrics which enable them to be distinguished with little difficulty from chemically deposited mosaics. The intergranular boundaries are irregular in form instead of plane; the grain shape is very variable; grain size varies much from place to place; and the distribution of grain size is notably patchy.

In limestones grain growth appears to affect only the very fine mosaics with grain diameters from 0.5–4μ. These include calcite-mudstone, the walls of foraminifera, algal frameworks, Bahamite particles, and ooliths.

ACKNOWLEDGMENTS

The author is indebted to his wife and to Phillip R. Brown for much helpful criticism of the manuscript.

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FIGS. 1–6, SEPTEMBER, 1959

PETROGRAPHY AND FACIES OF SOME UPPER VISÉAN (MISSISSIPPIAN) LIMESTONES IN NORTH WALES[1]

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ABSTRACT

The petrography of a dominantly limestone succession is interpreted in terms of three depositional environments—deep shelf (or upper slope), lagoonal, and shallow shelf. Special attention is paid to the composition of the limestones, their grain size and their depositional fabrics, all of which have been studied in thin section. Additional evidence is derived from a study of the associated shales and quartzose sandstones.

Point-counter analysis of the four main components in the limestones (calcite-mudstone, crinoidal debris, "other detritus," and calcite-siltstone) has led to a five-fold natural classification into (1) shelly calcite-mudstone, (2) shelly calcite-siltstone, (3) coquina-lutite, (4) bioclastic calcarenite, and (5) crinoidal calcarenite. The coquina-lutites are again divided into three types on a basis of grain size and planar shape orientation of the skeletal particles. The classification is further supported by the results of grain size analysis.

Types 1, 2 and 3 and the clay shales are interpreted as lagoonal, types 4 and 5 as deep-water shelf (or possibly upper slope), and the quartzose sandstones, calcareous sandstones, sandy limestones and silty shales as shallow water shelf deposits.

Possible fungal objects are recorded for the first time from the limestones.

INTRODUCTION

The purpose of this study was the separation, in a dominantly marine limestone succession, of the various carbonate facies on a basis of skeletal composition, grain size and depositional fabric, followed by an analysis of their depositional environments.

The region (fig. 1) was selected because the Viséan sediments there, though only moderately well exposed, include—besides limestones—shales and quartzose sandstones which yield additional evidence of environmental conditions. It extends from Castell, Denbighshire (Nat. Grid Ref. SJ/203574) in the south to Rhydymwyn, Flintshire (Nat. Grid. Ref. SJ/207668) in the north, a distance of about seven miles. The succession embraces the transition from the dominantly limestone facies of the Carboniferous Limestone (Viséan) to the sandstone-shale facies of the Millstone Grit (Namurian)—all within the Upper Mississippian (fig. 2). Attention was restricted almost entirely to the upper part of the Viséan succession *above* the Lower Brown Limestone. The total Viséan succession is about 3000 ft thick and crops out in a band about a mile and a half wide, striking north-south and dipping eastwards at about 20°.

FIG. 1.—Position of the area.

The conclusions in this paper are based on the results of lithological mapping on a scale of 1:10,560 and the examination of about 250 thin sections.

In a series of papers between 1876 and 1886, Morton divided the Lower Carboniferous Limestone succession of North Wales

[1] Manuscript received March 14, 1959.

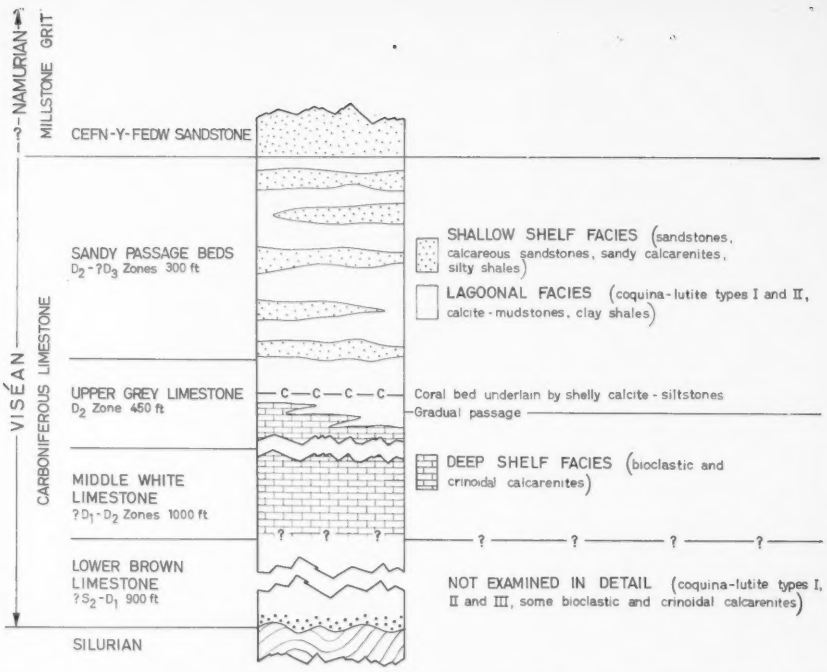

?-NAMURIAN
MILLSTONE GRIT
CEFN-Y-FEDW SANDSTONE
VISÉAN
CARBONIFEROUS LIMESTONE
SANDY PASSAGE BEDS
D2 - ?D3 Zones 300 ft
UPPER GREY LIMESTONE
D2 Zone 450 ft
MIDDLE WHITE LIMESTONE
?D1 - D2 Zones 1000 ft
LOWER BROWN LIMESTONE
?S2 - D1 900 ft
SILURIAN
SHALLOW SHELF FACIES (sandstones, calcareous sandstones, sandy calcarenites, silty shales)
LAGOONAL FACIES (coquina-lutite types I and II, calcite-mudstones, clay shales)
Coral bed underlain by shelly calcite-siltstones
Gradual passage
DEEP SHELF FACIES (bioclastic and crinoidal calcarenites)
NOT EXAMINED IN DETAIL (coquina-lutite types I, II and III, some bioclastic and crinoidal calcarenites)

FIG. 2.—Diagram of the Mississippian succession between Castell and Rhydymwyn, North Wales.

and its borders into three parts on a lithological and faunal basis (fig. 2).

Later the results of microscopical study of the limestones of neighbouring districts were published by T. A. Jones (1921) who for the first time showed the great importance of various micro-organisms in building these rocks. He believed that the limestones above and below the White Limestone were deposited "in quiet, shallow and partially land-locked waters," while the White Limestone represents "deeper, though still shallow sea conditions." His conclusions are to a great extent confirmed by the present study. A general summary of the palaeogeography is given by George (1958, p. 281–284).

CLASSIFICATION OF THE LIMESTONES

The three major components in the limestones are calcite-mudstone, crinoid debris and calcite-siltstone. For analytical purposes all other detrital material was combined into a fourth, composite, component called "other detritus." This consists of brachiopod and bryozoan debris, *Calcisphaera*, foraminifera, lamellibranch debris, broken shell fragments of about silt size, and rare quartz grains. Thin sections normal to the bedding plane were analysed by point-counter for the four components and the results were plotted on triangular diagrams (fig. 3).

Some of the calcite-mudstone recorded in fig. 3 is inside foraminiferal and other shell chambers or is present as mud aggregates (possibly faecal pellets or Bahaman-like sand grains—Illing, 1954).

On the basis of the analyses the limestones were divided into the following five main types:

Shelly calcite-mudstone[2]

[2] This rock is the familiar "calcite-mudstone" of stratigraphers. Unfortunately this term is also

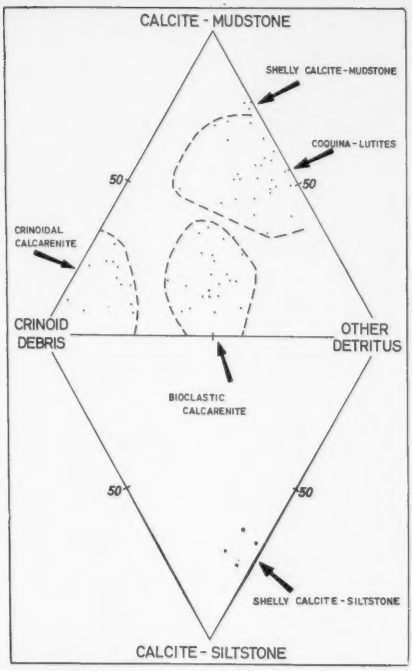


FIG. 3.—Point-counter analyses of thin sections of limestones.

Shelly calcite-siltstone
Coquina-lutite[3] (types I, II, III)
Bioclastic calcarenite
Crinoidal calcarenite

The coquina-lutites were further divided into three subtypes on the basis of particle orientation and grain size:

Type I.—Without preferred planar shape orientation of skeletal particles: coarse grained (figs. 4 & 6c).

Type II.—Skeletal particles with preferred planar shape orientation parallel to the bedding plane, and having roughly the same grain size as type I (figs. 4 and 6d).

Type III.—The finest-grained of the three (figs. 4 and 6e) with more calcite-mudstone, and with a preferred planar shape orientation of skeletal particles parallel to the bedding plane.

All seven limestone types are distinguishable in the hand specimen with the help of a hand lens.

The individualities of the three types of coquina-lutite and of the bioclastic and crinoidal calcarenites were confirmed by grain size analysis (figs. 4 and 5). Maximum apparent diameters (in thin section) were measured for 100 detrital particles, selected by point-counter, in each sample. Care was taken to exclude obvious cement including secondary crinoid rims, but it was not possible to exclude the cement in the calcite-mudstones and calcite-siltstones.

In all the limestones the non-carbonate residues, after digestion with hydrochloric acid, are: black opaque grains, chert, garnet, hematite, hornblende, ilmenite, leucoxene, pyrite, quartz and zircon.

LIMESTONES: CONDITIONS OF DEPOSITION

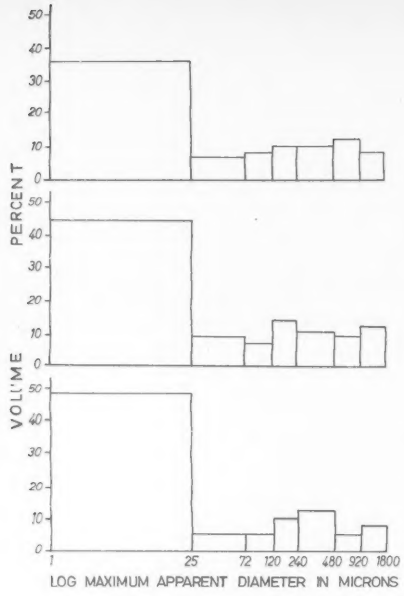
Shelly Calcite-mudstone

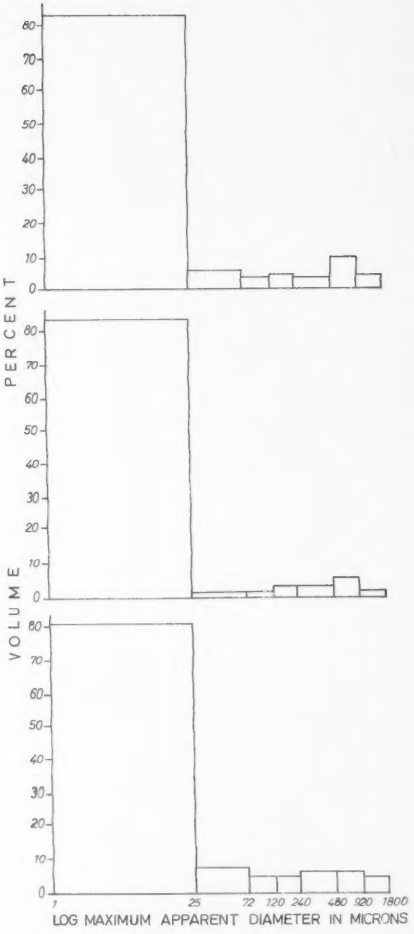
Appearance in thin section.—There is a little poorly sorted skeletal material, mainly unbroken lamellibranch valves and silt-sized skeletal fragments with no preferred planar shape orientation, in a matrix of roughly equant calcite grains with diameters from 0.5–4.0 μ (the calcite-mudstone proper). Crinoid debris and foraminifera are rare (figs. 3 and 6a).

Interpretation.—Fine-grained carbonate sediments are often explained as products of the comminuting effects of scavenging organisms. However, this explanation is not applicable to the North Wales calcite-mudstones because the grains in the matrix are more or less equant and the lamellibranch valves are commonly undamaged. It seems probable that the carbonate mud was precipitated, possibly biochemically, and the

used to describe the pure calcite-mudstone (grain size 0.5 to 4 μ) which can be recognized as a separate component in the limestones. It is felt that the latter use of the term is the more valuable and should be strictly adhered to. Consequently a limestone consisting dominantly of calcite-mudstone, but with some skeletal debris, is modified by the term "shelly."

[3] Coquina-lutite: the dominant component is skeletal debris of sand and silt grade which imparts to the rock a coarser texture than that of the shelly calcite-mudstone, though calcite-mudstone is present as matrix.





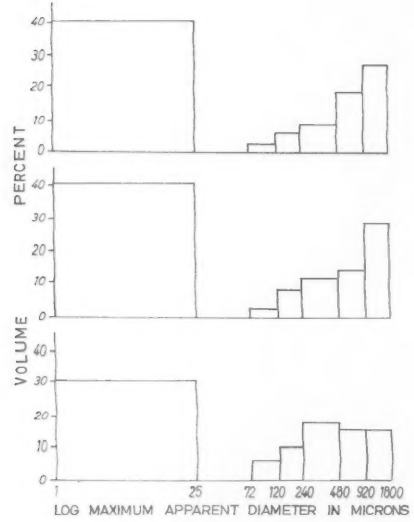
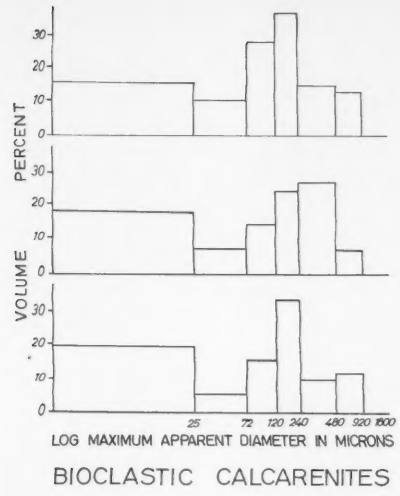
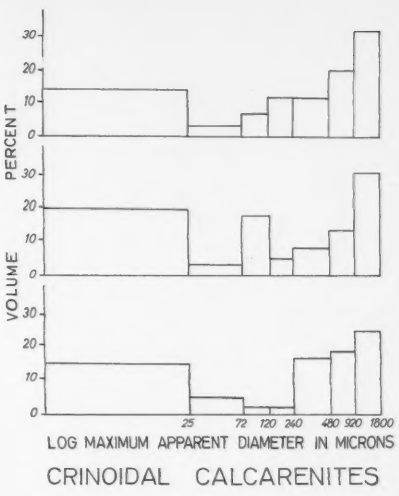


FIG. 4.—Point-counter analyses of grain size in nine samples of coquina-lutite.
As the limestones are monomineralic, volume is equivalent to weight.





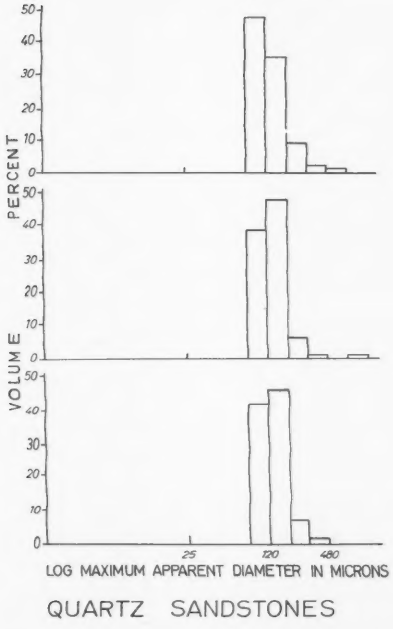


FIG. 5.—Point-counter analyses of grain size in three samples each of bioclastic calcarenite, crinoidal calcarenite and quartz sandstone.

scarcity of skeletal material may be an indication that life was impeded by high salinity.

The undamaged state of the lamellibranchs may point to sedimentation rapid enough to bury the shells before they could be broken by scavengers. The presence of mud indicates that bottom traction was not active during deposition.

Bottom traction includes rolling and sliding of particles along the bottom, and saltation. Only gravels, sands and the coarser silts can move in this way and their threshold velocities are such that the associated turbulence prevents the settling of finer particles. Thus, wherever mud is found, bottom traction must at least have been in abeyance while this settled from suspension. Where mud is more or less evenly distributed throughout a sediment bottom traction must have been inactive all the time.

The lack of any algal structures suggests that the muds are not the debris of an algal reef (compare algal reefs described by Wolfenden, 1958). However, until more is known about algal products it is not possible to say definitely whether the muds were of algal origin or not.

There are no needle-shaped crystals in these sediments, possibly an indication that there was no aragonite *at the time of cementation*—whatever may have been the composition of the mud before that time.

The shelly calcite-mudstones appear to be lagoonal sediments; the evidence for this comes partly from the associated clay shales which are discussed on a subsequent page.

Shelly Calcite-siltstone

Appearance in thin section.—This is composed of equant silt-sized calcite grains, with a range of diameters from about 5μ to about 35μ with little variation among samples, and a small amount of broken skeletal debris (figs. 3 and 6b). Unlike the shelly calcite-mudstone there is, in the shelly calcite-siltstone, little material intermediate in size between siltstone proper and the skeletal debris. This skeletal debris includes broken brachipods and bryozoans, crinoid debris and foraminifera. In some calcite-siltstones there are scattered quartz grains of silt or sand size. There is no preferred planar shape orientation of the skeletal particles.

This is the rock known locally as Aberdo limestone.

Interpretation.—The uniformity of grain size and the purity of the silt grade matrix suggest a chemical origin. This is supported by the scarcity of skeletal remains. The damaged state of the fossils and the absence of a preferred planar shape orientation may be the result of scavenging. The presence of quartz grains suggests slight bottom traction.

Coquina-lutites

Appearance in thin section.—These contain skeletal debris in a matrix of calcite-mudstone (figs. 3, 4, 6c–6e). The skeletal material includes crinoid debris, brachiopods (usually broken and angular), foraminifera, bryozoans, *Calcisphaera* and indeterminate skeletal remains of about silt size. Brachiopods are the most abundant. There is more pyrite in the coquina-lutites than in any other limestone type.

The features by which the three types of coquina-lutite are distinguished from each other have already been described.

Interpretation.—The presence of calcite-mudstone indicates that bottom traction was not active, so that the benthonic fauna (most of the foraminifera and brachiopods) must have been indigenous. The planktonic and nectonic fauna (other brachiopods, a few foraminifera and *Calcisphaera*) must have been carried in suspension. The small size and fragmentary nature of the crinoid debris suggest that this, too, was exotic.

In thin section all the brachiopod material looks very similar suggesting a restriction in species but abundance in individuals. If this is so, then the conditions of deposition may have been abnormal, for example lagoonal, brackish to estuarine, slightly hypersaline as suggested by Beales (1956, p. 862).

Reducing conditions *under* the sediment surface are indicated by the pyrite content. The pH of open sea water today varies generally between 7.5 and 8.5 and the Eh between 0.0 and 0.2. Krumbein and Garrels (1952, p. 26) have pointed out that pyrite and calcite can form and be stable in an environment in which the pH is approximately 8.0 and Eh approximately -0.3. It is, however, important to note that this Eh value does not necessarily imply a reducing

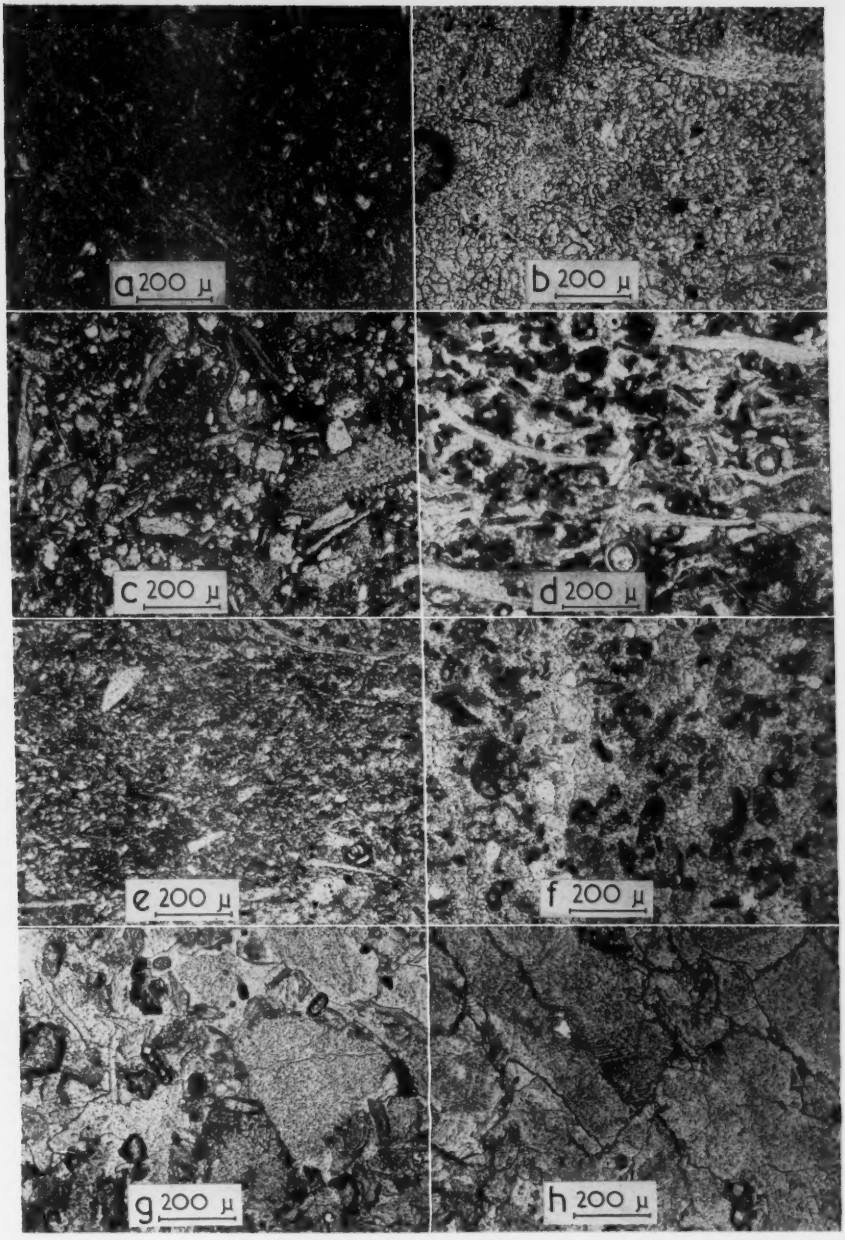


FIG. 6.—Photomicrographs of vertical sections through the seven limestone types. a. Shelly calcite-mudstone. b. Shelly calcite siltstone. c. Coquina-lutite type I. d. Coquina-lutite type II. e. Coquina-lutite type III. f. Bioclastic calcarenite. g. Bioclastic calcarenite. h. Crinoidal calcarenite.

marine water environment. Marine open circulation may exist down to the floor, while below this level there may be a reducing environment owing to an inadequate supply of oxygen.

The Viséan bottom waters are unlikely to have been so poorly aerated as to be reducing, since a benthonic fauna existed. Therefore, it is concluded that the pyrite formed after deposition.

From the close association of the coquina-lutites with black terrigenous muds it is concluded that they were deposited in a shallow, partly enclosed lagoon.

In coquina-lutite type I (fig. 6c) the absence of preferred planar shape orientation of the skeletons seems to indicate working of the sediment by scavengers. This was found to explain the absence of stratification in the bottom sediments of two Wisconsin lakes (Twenhofel and Broughton, 1939, p. 251; Twenhofel and McKelvey, 1939, p. 109), and was suggested for limestones (similar to coquina-lutites) of the Mississippian of the Ozark region of the United States by Moore (1957, p. 109).

Houbolt (1957) has described muddy skeletal calcarenites, containing unabraded fragmentary skeletal material, on the present floor of the Persian Gulf outside the surf zone. The skeletal material is mixed with calcite-mud. Houboult concludes that, since the skeletal material cannot have been broken by wave action, it must have been broken by scavengers, such as fish, worms, algae and molluscs. He supports this contention by reference to the observations of Darwin (1889), Murray and Lee (1909), Yonge (1930), and others.

Dapples (1942, p. 124) refers to instances where lamination is developed in modern sediments in spite of the action of scavengers and burrowers, simply because the rate of deposition exceeds the rate at which the organisms can turn over the sediment. This may have been so during the deposition of coquina-lutite types II and III which show good preferred orientations of skeletal particles.

Among scavengers, worms probably played an important part, as they thrive today in great numbers in shallow sandy and muddy bottoms (Dapples 1938, p. 58). According to Davison (1891) they can bring to the surface at least 1911 tons of material per year per acre.

A depth of water of about 9–12 m during the time of deposition is indicated by the presence of *Productus giganteus*, judging by the occurrence of similar types today (Dr. J. W. Jackson, personal communication). Similar thick-shelled brachiopods are found today in shallow parts of the Pacific and Indian Oceans, especially on coral shelves, and it is, perhaps, significant that in some of the Viséan limestones *P. giganteus* occurs interbedded with coral debris.

Coquina-lutite type II (fig. 6d) has a preferred planar shape orientation of skeletal debris showing absence of sub-surface scavenging and burrowing. Yet most of the skeletal remains are broken. It is concluded, therefore, that in the absence of bottom traction they were broken by scavengers *before* burial.

Coquina-lutite type III (fig. 6e) also has a preferred orientation. This sediment was probably deposited under much the same conditions as type II, but was either exposed to greater attack by scavengers and burrowers before burial, thus attaining a finer particle size, or was deposited from suspension in quieter water.

The differences in the grain size analyses of the three types presumably reflect also other undetermined environmental factors.

Bioclastic Calcarenite

Appearance in thin section.—This consists of skeletal debris in calcite cement (figs. 3, 5, 6f, 6g). Some specimens contain a few scattered quartz sand grains (more than in the other limestones), others a few ooliths.

The skeletal remains consist of crinoid debris, brachiopods and bryozoans (usually broken), *Calcisphaera*, foraminifera (usually entire), algal fragments and indeterminate debris of about silt size. Sorting is poor and the skeletal fragments are angular. The finer examples contain patches of calcite-mudstone or calcite-siltstone as matrix. There is no preferred planar shape orientation of skeletal particles.

In the field these limestones are in places current-bedded and show flute-casts. Locally they contain coral debris.

Interpretation.—The activity of bottom traction is indicated by the quartz sand

grains, occasional ooliths, current-bedding, and rare flute-casts in underlying clay shale. The currents were, however, so weak that sorting was inefficient and abrasion negligible.

The absence of a preferred planar shape orientation of the skeletal particles, and their broken nature, are presumably the results of scavenging.

Most of the larger skeletal particles are crinoids. These had a high porosity and so were lighter than the other debris and would have been moved simultaneously with smaller shell fragments.

A shelf sea appears to be the most probable environment of deposition. Upper slope conditions seem possible but less likely. The sediment is almost pure carbonate and modern deposits of similar purity are known on the Bahama Banks and in the Persian Gulf on floors less than 20 m and 60 m deep respectively. Carbonate-rich sediments are known from the Paria-Trinidad-Guiana shelf and the upper slope (Koldewijn, 1958; Nota, 1958), but these contain much terrigenous material. In the Mississippian calcarenites, the calcareous algae, ooliths and corals indicate a shelf source for some of the sediment. The poor sorting and angularity are reminiscent of the "skeletal calcarenites" of Houbolt (1958) in the Persian Gulf. Flute-casts are known from a number of marine and terrestrial environments and offer no clear indication. Although the particles have undergone bottom traction they are not necessarily exotic since the movement may have been oscillatory.

The sizes of the foraminifera vary directly with the sizes of the other detrital particles, with appropriate allowance for the large crinoids. It follows that this relationship is a result of sorting by currents, as proposed by Carter (1951), and that the foraminifera are a death assemblage.

The rare small patches of calcite-mudstone do not indicate the continual depositional process. The mud can only have arrived in suspension, to filter down into the pores of the skeletal sand, during periods of reduced turbulence.

Some of the skeletal remains have a uniformly thin coating, about 10 μ thick, of calcite-mudstone. The absence of a concentric pattern within these mudstone envelopes makes it unlikely that they are oolitic. They may be algal coatings deposited on the skeletons before final deposition. This again would imply shallow water.

Crinoidal Calcarenite

Appearance in thin section.—Some of these limestones (figs. 3, 5, 6h) consist almost entirely of cemented crinoid debris with a little indeterminate fragmental skeletal material. In others there are foraminifera and broken remains of bryozoans and brachiopods. All are cemented by calcite. There is no preferred planar shape orientation of the skeletal material. Scattered patches of calcite-mudstone occur interstitially.

Interpretation.—The high crinoid content, generally unabraded material, suggests that the Viséan crinoids were as gregarious as their modern successors (Shrock and Twenhofel, 1953; p. 669, Laudon, 1957). The other crinoid-bearing calcarenites show transported crinoid debris, as in the bioclastic calcarenites, or extremely small and fragmentary debris of doubtful source, as in the coquina-lutites. The distinct gap between the crinoidal and the bioclastic calcarenites in fig. 3 shows that there is no gradation between the two.

It is generally believed that both the modern and Mississippian stem-bearing crinoids are sessile benthonic. Living specimens have been found on all sorts of sea bottoms—muddy, sandy or rocky—at depths from low tide mark to about 10,000 m. They live in depressions and in places more or less protected from big waves, apparently preferring situations undisturbed by bottom traction.

The absence of preferred planar shape orientation may indicate working of the sediment by scavengers, though it is more probably a result of the absence of markedly discoid or elongate particles.

The presence of calcite-mudstone indicates that bottom traction was slight and supports the view that the crinoids grew where they are now found.

A Note on Possible Fungal Remains

When the limestones are digested with hydrochloric acid some dust-like particles (100–150μ diameter) float on the surface of the liquid, despite the care taken to avoid

contamination by modern dust. They are brownish, roughly spherical, and have a wrinkled appearance. After removal, washing in distilled water, and drying, the particles just sink in a mixture of 75 percent alcohol and 25 percent water.

These are organic remains and are known to occur in some calcareous rocks and shales with a high calcium carbonate content; in particular they are present in the calcareous rocks of the Scottish Oil Shale Group (Professor L. R. Moore, personal communication).

Professor Moore has been good enough to report in detail on the particles. He states that they are organic, but not spores in the accepted sense. They could be fungal resting spores since they carry evidence of fungal hyphae; on the other hand their refractive index and general appearance do not substantiate this.

Professor Moore favours the idea that the bodies may be compound, that some original object may form the basis, as for example an algal body or a related type such as *Cloecapsomorpha* or a *Calcisphaera*. It is possible that the ornamentation may result from the presence and attack of minute parasitic fungal bodies.

TERRIGENOUS SEDIMENTS: CONDITIONS OF DEPOSITION

Clay Shale

Appearance in thin section.—This is a mixture of mud grains (by definition less than 5μ) and silt grains (by definition 5 to 70μ). Texturally the shale is a silty mud. The mud and silt form impersistent laminae generally less than 2 mm thick. The grains consist of black material (probably carbonaceous), clay minerals, quartz, mica, chert, leucoxene, and pyrite, with rare skeletal debris.

In the field the clay shales lie mainly in the Sandy Passage Beds and are interbedded with coquina-lutites, the contacts being sharp. The shales are mainly black, but sometimes grey or brown. The grains less than 1.5μ are almost all black and opaque. As the clay shales have been shown by chemical analysis to contain roughly 30 percent of organic matter the small grains are probably mainly carbonaceous.

A single analysis has yielded a non-carbonate residue of 97 percent by weight. Of this residue 70 percent consists of black opaque grains.

Interpretation.—The high organic content of these clay shales and the presence of much authigenic pyrite indicate that reducing conditions must have prevailed at some time.

The preservation of laminae and a preferred planar shape orientation of the skeletal debris indicate the absence of subsurface scavengers. The scarcity of skeletal remains points also to the paucity of benthonic life. Thus is can be concluded that where these sediments were deposited there was inadequate oxygen for aerobic life, not only in the pore water of the sediment, but also in the bottom sea water. In other words deposition took place in stagnant, anaerobic, reducing conditions. Richly organic deposits (up to 25 or even 30 per cent of organic material) are typical of modern ill-ventilated water bodies (Kuenen 1950, p. 13).

These conditions are known to be typical of poorly oxygenated basins. For example, below about 2 cm from the sediment surface in Lake Windermere, Westmorland, the richly organic muds have a negative Eh all the year round (Mortimer 1949, p. 355). In the Orinoco deltaic estuarine and pro-delta facies of the Gulf of Paria (van Andel and Postma, 1954, p. 165) grey-black muds (p. 111) have a pH from 6.4 to 7.5, a negative Eh (p. 96, 97), abundant pyrite (p. 161) and an impoverished or a non-existent marine fauna (p. 119–122).

The fine grain size of the Viséan clay shales show that they were pelagic.

Silty Shale

Appearance in thin section.—The silty shales are red and grey, with no pyrite, and have a non-carbonate residue of approximately 50 percent by weight. This includes quartz, mica, chert and felspars. Texturally they are muddy siltstones. Most of the calcite is in the form of calcite-mudstone. The red colour indicates iron oxide and the iron may be present in the grey silty shales in the reduced state. The red colour is particularly strong and uniform in the finer patches.

In the field these sediments occur high in the succession interbedded with coquina-

lutites and calcareous sandstones, commonly lying just above and below the quartzose sandstones.

Interpretation.—The silty shales did not undergo the reducing conditions that affected the clay shales. The iron oxide was not supplied by percolating water from overlying beds and so presumably arrived with the sediment. As the amount of dissolved iron in modern open sea water is very low, and as iron will not precipitate at such low concentrations, it is likely that the iron arrived as a colloidal suspension. This receives some support from the density of the red colour throughout the *finest-grained* parts of the shale. Scattered lenticles of reddening in the underlying grey variety are restricted to coarse laminae and are probably the result of downward percolation of iron solutions from the red shales into the larger pores.

The silty shales appear to have been deposited in well-aerated water. The calcite-mud was most likely chemically precipitated, in the same way as that in the shelly calcite-mudstones.

Quartzose Sandstones

Appearance in thin section.—Quartz grains are combined in various proportions with the components of bioclastic calcarenite and with ooliths, giving a range from pure quartz sandstone to calcareous sandstone and sandy calcarenite. The largest quartz grains have intermediate diameters of about 4 cm. There are few grains below 25μ (maximum apparent diameter), and about 95 percent (by volume) of the grains lie between 150–650μ. The cements are quartz and calcite.

Detailed comparison of the pure quartz sandstones with those of the overlying Millstone Grit facies has revealed their close petrographic similarity.

In the field there are current-bedding units up to 10 ft thick.

Current-direction analysis, by measurement of the azimuthal orientation mainly of apparent longest grain axes in thin section and also of the direction of dip of the slip-off slopes of current-bedding, has yielded *axes* of water movement for the pure quartz sandstones which lie almost entirely within the arc WSW-SSW, over the whole area. There is no clearly dominant *direction* of movement.

Interpretation.—The presence of current-bedding, quartz pebbles, and ooliths shows the importance of bottom traction during deposition. The quartz must be exotic, so that it is probable that the skeletal particles, too, may be many miles from their sources.

The constancy of the axes of current movement may be a product of tidal flow or —perhaps more probably—of a prevailing wind.

GENERAL DEPOSITIONAL CONDITIONS

The main conclusions drawn from the preceding laboratory studies are summarised in table 1.

It is now possible to consider the field relationships of the different types of sediment and their implications.

The Tropical Climate

According to Rodgers (1957, p. 4) modern shelf carbonate deposits lie almost entirely within 30°N and S of the Equator. It is, therefore, probable that similar deposits of the Viséan were laid down within roughly similar limits, allowing for some extention in a possibly warmer world climate. This is borne out by evidence from rock magnetism which indicates a Carboniferous North Pole about 50° N and 140° E (data in Creer, Irving and Runcorn, 1957) and gives Britain a latitude well within the tropics.

The Three Facies

In the field the various sediments occur in three major associations (fig. 2). The lower part of the succession consists of bioclastic calcarenites with subordinate crinoidal calcarenites. There is an interbedded passage between these and the overlying association of sediments. This higher association comprises mainly coquina-lutites (types I and II) with subordinate shelly calcite-siltstones, shelly calcite-mudstones and clay shales. Higher in the succession these fine sediments are interbedded with a third association consisting of quartz sandstones, calcareous sandstones, sandy calcarenites, and silty shales.

The three associations of sediments appear to represent quite distinct depositional environments. These are believed to be, in

TABLE 1.—*Summary of main conclusions based on laboratory examination*

Rock type	Mode of transport	Depositional environment
Shelly calcite-mudstone	Suspension (chemical precipitate)	Quiet sea, too saline for abundant animal life. No bottom traction
Coquina-lutites	Mud in suspension. Skeletons mainly indigenous benthonic	Sea quiet. Water oxygenated enough for benthonic life. Bottom scavengers or burrowers only active during deposition of type I. No bottom traction
Clay shale	Suspension	Quiet, sheltered water, with insufficient oxygen to support life other than bacteria, etc. Absence of bottom traction
Bioclastic calcarenite	Bottom traction	Open sea, good ventilation by turbulent water
Crinoidal calcarenite	Little movement, skeletons indigenous benthonic	Sheltered well oxygenated sea floor undisturbed by bottom traction
Silty shale	Bottom traction of detrital grains. Possible chemical precipitation of calcite-mud	Slight bottom traction. Probable inflow of fresh iron-bearing water. Aerated bottom
Quartzose sandstones	Bottom traction	Open sea, good ventilation by turbulent water

ascending order, deep shelf (possibly upper slope), lagoonal and shallow shelf.

Deep shelf facies.—This facies is characterised by bioclastic calcarenites consisting of skeletal debris which has undergone only slight bottom traction. Small amounts of interstitial calcite-mudstone testify to periods of reduced turbulence. Crinoid meadows flourished in some sheltered localities.

The bioclastic debris was deposited in deeper water than the materials of the sandy limestone in the shallow shelf facies. This is apparent from its lower degree of abrasion and poorer sorting, the rarity of current-bedding and ooliths, and the presence of interstitial calcite-mudstone.

The bioclastic calcarenites compare favourably with the modern "skeletal calcarenites" of the central part of the Persian Gulf (Houbolt, 1958, p. 46, 53, 87). These sediments are less abraded and less well sorted than the shallower "rounded calcarenites" which lie within the zone of breakers. They have a small silt content ($<53\,\mu$). The Viséan bioclastic calcarenites are rather finer-grained (maximum of $920\,\mu$) than Houbolt's skeletal calcarenites (maximum 2 mm) and show, on the whole, less signs of abrasion and of breakage of larger shells.

Houbolt's skeletal calcarenites are found beyond the zone of breakers, at a depth of 18 m or more, where transport by waves is still active and where grains up to $450\,\mu$ are at times brought into suspension. During quieter periods fine grains ($<53\,\mu$) are deposited and trapped between the larger particles. Although the skeletal debris is moved about locally it shows little rounding and the angular broken fragments appear to be a product of scavenging activity (Houbolt, 1958, p. 46) rather than bottom traction (not uncommonly broken shell debris is embedded in silt). Thus the sediment has not been transported from the breaker zone, but is autochthonous, though capable of local movement.

It seems probable that the Viséan bioclastic calcarenites were also autochthonous and were affected only by oscillatory wave motion. Such a local origin may acacount for the rarity of quartz which is so plentiful in the younger, shallow shelf sediments.

The lagoonal facies.—The sediments of this facies include coquina-lutites (types I and II), shelly calcite-mudstones, shelly calcite-siltstones, and clay shales. They were deposited on a floor where bottom traction was negligible. The clay shales, deposited in poorly ventilated water isolated from the

open sea, are interbedded with carbonate sediments of local origin. The alternation is conveniently examined by considering first the clay shales.

In the shallow Viséan tropical seas, the small range of climatic variation, coupled with the presence of a barrier against the influence of open sea water, must have made possible the development of bodies of relatively stagnant ill-ventilated water. It seems that mud (clay shale) was deposited in these conditions on a floor practically devoid of benthonic life. Organic material settled from suspension and underwent anaerobic decomposition giving rise to abundant pyrite.

At times the water must have become better oxygenated, perhaps as a result of flooding from the open sea. This would have enabled benthonic life to grow. At the same time, evaporation from the surface of the shallow water—possibly assisted by organic activity—appears to have caused the precipitation of carbonate mud. While the coquina-lutites were being deposited the degree of oxidation of the bottom water probably fluctuated causing variations in the activity of scavengers below the sediment surface. Where these were active coquina-lutite type I was deposited.

Whenever these enclosed waters became too saline for animal life and precipitation of calcium carbonate greatly exceeded the supply of organic and terrigenous mud (clay shale) then calcite-mud with a low skeletal content accumulated.

While the implication here would appear to be that the rate of precipitation of carbonate mud varied with time, it is not clear whether the supply of organic and terrigenous suspended matter fluctuated or whether it was constant and at times swamped by carbonate precipitation.

Taken together these depositional conditions seem to be characteristic of a lagoon. Although a barrier must have separated the lagoon from the open sea there is no evidence for algal or coral reefs, though a bed of overturned and somewhat broken corals (mainly colonial *Lithostrotion* and *Dibunophyllum*) can be traced near the top of the Upper Grey Limestone. Isopachs for the D zone indicate a coast running NW-SE less than twenty miles to the southwest (George, 1958, p. 284).

Shallow shelf facies.—These sediments, mainly quartz sandstones with varying admixture of bioclastic debris, were deposited in shallower water than the bioclastic calcarenites. The skeletal particles are more rounded and better sorted and include ooliths. The sands are commonly current-bedded. Beds of silty shale, immediately below and above the quartzose sandstones, indicate less turbulent periods.

These sediments are finer-grained than the rounded calcarenites of the Persian Gulf, which were formed within the breaker zone. Thus it seems possible that these Viséan sediments were deposited outside that zone, though the quartz, being terrigenous, must have passed through it. The abundance of quartz in these sediments and its scarcity in the earlier bioclastic calcarenites may imply the inability of wave movement to carry quartz to the greater depths where bioclastic calcarenites were forming, rather than the late appearance of a source of quartz sand.

The change from Carboniferous Limestone facies to Millstone Grit facies.—The nature of the passage from marine carbonate deposition to the marine sand and mud deposition of the Millstone Grit facies can now be summarised. Shoaling of the sea was accompanied by an increased supply of terrigenous sand arriving by bottom traction. The change was not gradual, but spasmodic, and was interrupted by occasional retrogressions. Deeper shelf or, perhaps, upper slope, conditions began to alternate with, and then gave place to, lagoonal conditions. Later, shallow shelf conditions alternated with, and finally replaced, the lagoonal.

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FIGS. 1–5, SEPTEMBER, 1959

SIZE AND SHAPE OF ROCK-FRAGMENTS IN TUSCARORA SCREE, FISHING CREEK, LAMAR, CENTRAL PENNSYLVANIA[1]

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ABSTRACT

Variation in size, size-sorting, and axial proportions of fragments from a Tuscarora Scree in Central Pennsylvania is similar to variation in pebbles and quartz grains from other sedimentary deposits. The largest source of variation occurs among fragments and no pattern of variation is evident over the small area sampled. This is in accord with expectations on the basis of sampling sediments which lie near their source and have suffered little selective sorting.

INTRODUCTION

One of the earliest formed sedimentary deposits, occurring close to source, is represented by talus and scree; modification in this case is minor and the detritus represents the first stage in the formation of the final sedimentary rock. It would be of some interest to learn the magnitude of variation in size and shape of such material as some idea of what a source material is like.

Scree and talus represent the first, or at least a very early, stage in dispersal of detritus and should, therefore, belong in stage I of the model proposed by Griffiths (1955; and, in press); in other words, very little selective sorting has taken place and the deposit should, in a first approximation, be, homogeneous in size and shape. There should be no, or very little, layering of the deposit; and therefore it can be predicted that variance within a sample among fragments will be large, whereas between sample means variation will be near zero. The present experiment is a reconnaissance investigation of the problem and an attempt to decide what a comprehensive test of this prediction would entail.

SAMPLING AND EXPERIMENTAL DESIGN

The location chosen for examination was a scree of boulders from the Tuscarora Formation; the fragments consist of quartzite and are uniform in composition so that the size and shape variation is not affected by any major change in composition.

The deposit is 3 miles east of Lamar on State route 780 along Fishing Creek, in Central Pennsylvania. Field examination of the deposit suggests that it is made up of boulder trains which are roughly triangular in shape and with their apices uphill. No attempt was made at this stage to differentiate boulder trains; furthermore, while it appears likely that there is a size and shape gradient from higher elevation to lower, this was not tested for in the present experiment. The main objective was to find the magnitude of variation and decide how the deposit could be most efficiently sampled.

Two lines, 10 paces (1 pace = approximately $2\frac{1}{2}$ feet) apart, were drawn across the exposed surface of the deposit perpendicular to the axis of the triangle and four operators, Chen (C), Huh (H), Spalding (S), and Wood (W) were assigned to alternate spots. The first spot was selected at random and the subsequent spots were uniformly spaced 10 paces apart. At each spot each of two operators measured four contiguous fragments.

This sampling net results in a completely randomized design for the deposit; but as each line is common to all operators, the operators are crossed with lines. The field arrangement is given in figure 1.

The experimental design is summarized in table 1 in the form of and analysis of variance. The sources of variation show how the variance is apportioned to the different effects: in this design the blocking, or stratification, of operator-pairs is intended to remove the effect of operator difference; and if the interaction term operator times lines is small, then the design will have succeeded

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FIG. 1.—Field arrangement of operators in lines on Tuscarora Scree.

Ops.	CH	SW	CH	SW	
Line 1	8	8	8	8	No. of fragments
Ops.	CH	SW	CH	SW	
Line 2	8	8	8	8	No. of fragments

because differences among operators is orthogonal to differences among lines.

Since each operator measured four fragments at each spot, a total of 64 fragments was measured, 16 per operator, 32 per line. Each set of 8 is treated as an operator-pair unit.

In a second stage of the experiment a single operator (JCG) measured four contiguous fragments at each of six locations selected at random outside but next to the preceding area. This sampling set led to a completely randomized design summarized in table 2.

This part of the experiment should yield a result closely similar to the first if the deposit is homogeneous.

TECHNIQUE AND PROCEDURE

Three measurements were performed on each fragment: the longest axis, "a"; the intermediate axis, "b"; and the shortest axis, "c". The "b" axis is the shortest axis in the plane of the maximum projection; the "a" axis is defined as the longest axis in the same plane perpendicular to "b"; the "c" axis is the longest axis perpendicular to the "a–b" plane. The selection of axes was simplified by the rectangular shape of most of the fragments. One difficulty arose because the fragments were too large to be moved; it was sometimes difficult to measure the short axis because the fragment was partly covered by its neighbors.

The measurements made in the field in feet and inches were recorded on data cards, and these data were transformed to millimeters and then to the phi scale (Krumbein, 1936).[1] The subsequent analysis was performed on the data expressed in phi units.

DESCRIPTION OF THE DATA

The data comprise measurements of 3 axes on 64 and 24 fragments of scree from the two experimental designs. The frequency distributions of all 88 measurements are expressed as histograms in figure 2a, b, c. The description of the data is best summarized by the descriptive statistics of table 3.

Now it is obvious that the mean sizes indicate coarse-grained fragments; however, the spread of size in each axis measured by the standard deviation ($s\phi$) is of the same

[1] A comprehensive table is now available which simplifies the phi transformation (Griffiths and McIntyre, 1958).

TABLE 1.—*Experimental design for analysis of variance of axis measurements of fragments from Tuscarora Scree, Fishing Creek, Pennsylvania*

No. of Items	Source of Variation	Degrees of Freedom	Expected Mean Square
$p=2$	Operator Pairs (p)	$(p-1)=1$	$\sigma_e^2+f\sigma_q^2+fq\sigma_{pl}^2+lfq\sigma_p^2$
$l=2$	Lines l	$(l-1)=1$	$\sigma_e^2+f\sigma_q^2+fq\sigma_{pl}^2+pfq\sigma_l^2$
	Operators by Lines	$(p-1)(l-1)=1$	$\sigma_e^2+f\sigma_q^2+fq\sigma_{pl}^2$
$q=4$	Spots within operators and lines	$pl(q-1)=12$	$\sigma_e^2+f\sigma_q^2$
$f=4$	Fragments within all classes	$plq(f-1)=48$	σ_e^2
$plqf=64$	Total	$(plqf-1)=63$	

TABLE 2.—*Second sampling plan across Tuscarora Scree at Fishing Creek, Central Pennsylvania*

No. of Items	Source of Variation	Degrees of Freedom	Expected Mean Square
$q=6$	Among Locations	$(q-1)=5$	$\sigma_e^2+4\sigma_q^2$
$f=4$	Fragments within Locations	$q(f-1)=18$	σ_e^2
$qf=24$	Total	$(qf-1)=23$	σ_{tot}^2

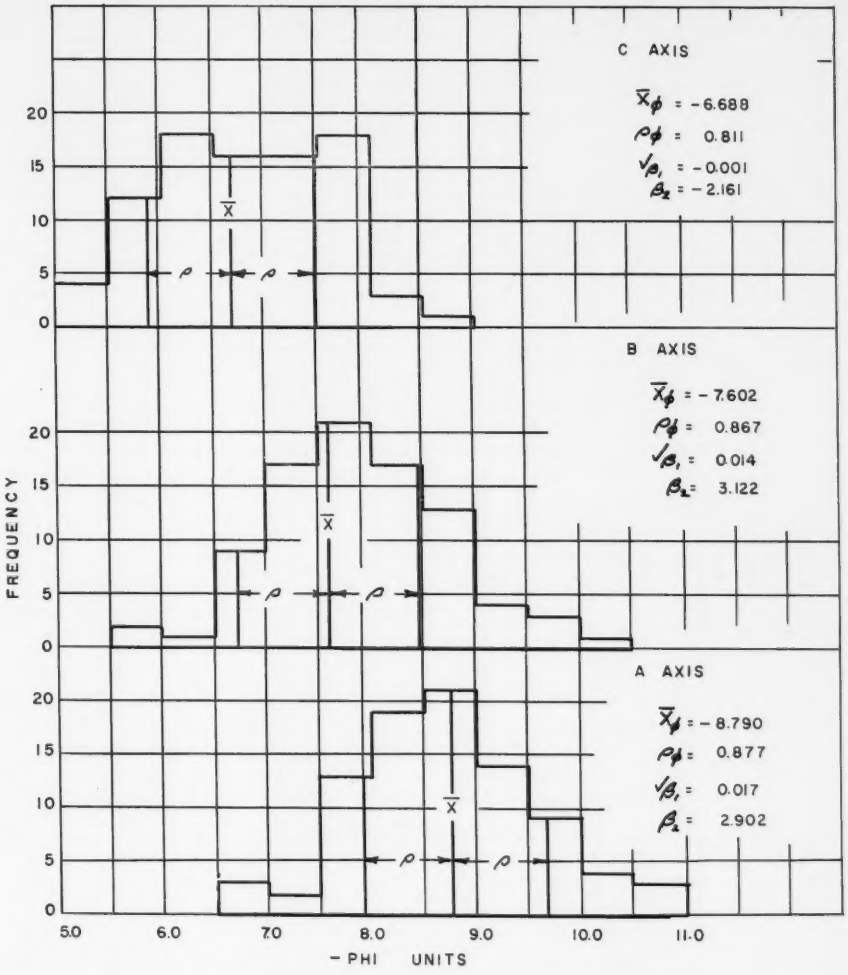

FIG. 2.—Frequency distributions of axial lengths in fragments from Tuscarora Scree: (a) a-axis, (b) b-axis, (c) c-axis.

order as that found in the sand range (i.e., this deposit is as well sorted as the average sedimentary rock). If subsequent investigations bear this out, then it may be deduced that the standard deviation, and more particularly the variance, is independent of the mean. This is a characteristic of random samples drawn from a normal population and is of considerable value in establishing both the population frequency distribution and the randomness of samples. These results are very different from those obtained on the basis of sieving and pipette sedimentation (Griffiths, 1951a, 1951b, 1952).

TABLE 3.—*Summary statistics for size measurements of fragments from Tuscarora Scree*

	Axis					
	a		b		c	
	No. of Operators					
	4	5	4	5	4	5
$\bar{X}\phi$	−8.781	−8.790	−7.609	−7.602	−6.719	−6.688
$s^2\phi$	0.671	0.768	0.761	0.751	0.702	0.658
$s\phi$	0.819	0.877	0.873	0.867	0.838	0.811
$\sqrt{\beta_1}$	0.079	0.017	0.004	0.014	0.001	−0.001
β_2	3.214	2.902	3.448	3.123	2.178	2.161
Sk	0.040	0.009	0.002	0.007	0.0005	−0.0005
K	0.214	−0.098	0.448	0.123	−0.178	−0.839
N	64	88	64	88	64	88

The asymmetry ($\sqrt{\beta_1}$) of these small samples is not likely to be a very stable estimator of the asymmetry of the population from which the sample of 88 was drawn, but as far as these data are concerned the skewness of the distribution is negligible. Similarly the kurtosis (β_2) is not significant although it tends to fluctuate quite widely; actually, the smallest sample size listed in the Biometrika Tables (Pearson and Hartley, 1954, p. 184, table 34) is n = 200, and 88 is considered too small a sample for testing. To all intent then, on the basis of these statistics, there is no reason to suppose that the population from which this sample of 88 was drawn is other than normally distributed, a feature of some import in pursuing the subsequent analysis.

It is also of some interest to note that the variances based on the sample of 64 from the first experimental design do not differ significantly (table 4) from the variances of the combined experiments, which is some justification for combining the results.

UNIVARIATE ANALYSIS: SIZE AND SORTING

The first step in analysis of the data is to compare means at the different levels of the experiment; the analysis of each of the three axes using the sampling design of table 1 are summarized in tables 5, 6, and 7 respectively. Here we are considering that the two lines are random samples of an infinite population of such lines and so are the operators, spots within lines, and fragments within spots, i.e., the infinite population model (Eisenhart, 1947; Bennett and Franklin, 1954, p. 348). Table 5 may be used as an example.

The basic error term in this analysis is the variation among fragments which is, more correctly, variation from all unassigned sources; this "error" is compared with variation among spots, and as the latter is the smaller, its contribution to the variance of the population is zero; in other words, the variation among spots is not significantly greater than error. The error may now be compared with the interaction term operators×lines and again this source is not significantly larger. In effect then, differences among spots and differences among operators×lines' means are not significant and these separate estimates of variance may be

TABLE 4.—*Variance ratios testing the differences in variability from the first and combined sampling plans*

	a axis	b axis	c axis
Variance Ratio	0.7683/0.6709 = 1.145	0.7615/0.7509 = 1.014	0.7022/0.6580 = 1.067
d.f.	88:64	64:88	64:88
F_{05}	1.49	1.51	1.51

TABLE 5.—*Analysis of variance of long "a" axis of Tuscarora Scree, Fishing Creek, Pennsylvania*
Data coded; observed "a" axis $= X_1'$; and $X_1 = -10^2(6+X_1')$

Source of Variation	Degrees of Freedom	Sums of Squares	Mean Square	F
Operator Pairs (p)	$(p-1)=1$	3,052.56	3,052.56	<1
Lines (l)	$(l-1)=1$	22,126.56	22,126.56	3.402
Operators by lines	$(p-1)(l-1)=1$	4,489.005	4,489.005	<1
Spots within operators and lines	$pl(q-1)=12$	49,940.375	4,161.698	<1
Fragments within all classes	$plq(f-1)=48$	312,160.500	6,503.34	
Total	$(plqf-1)=63$	391,769.00	6,218.556	
Pooled Error	61	366,589.880	6,009.670	

Coded $\bar{X}_G = 270.625$ Decoded $\bar{X}_G = -8.706$
Pooled $s_e = 77.52$ Decoded $s_e = 0.7752$

TABLE 6.—*Analysis of variance of intermediate "b" axis of Tuscarora Scree, Fishing Creek, Pennsylvania*
Data coded; observed "b" axis $= X_1'$; and $X_1 = -10^2(6+X_2')$

Source of Variation	Degrees of Freedom	Sums of Squares	Mean Square	F
Operator Pairs (p)	1	2,795.77	2,795.77	<1
Lines (l)	1	21,133.89	21,133.89	2.666 2.473
Operators by lines	1	178.89	178.89	<1
Spots within operators and lines	12	102,568.56	8,547.38	1.077
Fragments within all classes	48	380,884.25	7,935.09	
Total	63	507,561.36	8,056.53	
Pooled Error	61	483,631.70	7,928.38	

Coded $\bar{X}_G = 189.703$ Decoded $\bar{X}_G = -7.897$
Pooled $s_e = 89.76$ Decoded $s_e = 0.8976$

TABLE 7.—*Analysis of variance of short "c" axis of Tuscarora Scree, Fishing Creek, Pennsylvania*
Data coded; observed "c" axis $= X_1'$; and $X_1 = 10^2(5+X_3')$

Source of Variation	Degrees of Freedom	Sums of Squares	Mean Square	F
Operator Pairs (p)	1	17,523.15	17,523.15	2.490
Lines (l)	1	16,802.65	16,802.65	2.388
Operator by lines	1	1,269.13	1,269.13	<1
Spots within operators and lines	12	84,444.56	7,037.05	1.08
Fragments within all classes	48	312,806.25	6,516.80	
Total	63	432,845.74	6,870.57	

Coded $\bar{X}_G = 182.141$ Decoded $\bar{X}_G = 0.8289$
Coded $s_e = 82.89$ Decoded $s_e = -6.821$

considered separate estimates of experimental error. They are, therefore, pooled and the pooled error may be used to test the two main effects, differences among lines and differences among operators. The latter is obviously not significant and so may be ignored. This clearly shows that differences among operators measuring long axes are not significantly greater than differences among fragments. The differences among line means is some 3 to 4 times larger than the variation among fragments but this is not significant.

It may be concluded that the subdivision into lines and spots has not reduced the variation and hence the variation among

TABLE 8a.—*Analysis of variance of "a" axis, second sampling design*

Source of Variation	Degrees of Freedom	Sum of Squares	Mean Square	Variance Ratio (F)
Among sets	$(m-1)=5$	31,142.33	6,228.47	0.565
Within sets	$m(p-1)=18$	198,483.00	11,026.83	
Total	$(mp-1)=23$	229,625.33	9,983.71	

TABLE 8b.—*Analysis of variance of "b" axis, second sampling design*

Source of Variation	Degrees of Freedom	Sum of Squares	Mean Square	Variance Ratio (F)
Among sets	5	31,690.38	6,338.07	<1
Within sets	18	142,552.25	7,919.57	
Total	23	174,242.63	7,575.77	

TABLE 8c.—*Analysis of variance of "c" axis, second sampling design*

Source of Variation	Degrees of Freedom	Sum of Squares	Mean Square	Variance Ratio (F)
Among sets	5	16,870.33	3,374.06	<1
Within sets	18	112,399.00	6,244.39	
Total	23	129,269.33	5,620.41	

fragments is homogeneous. Similarly, differences among operators and operators×lines' inconsistency is negligible. Hence, in terms of the long "a" axis measurements, these data are homogenous. The same is true of the variation among the "b" axis measurements which are also homogeneous over the deposit sampled.

The "c" axis yields one item which is different, namely the contribution from differences among operator means is much larger than in the earlier experiments and compares in magnitude with differences among line means. The interaction is negligible so that operator differences do not affect the results, but why operators should vary much more in measurement of "c" axes than "a" or "b" axes is not clear; perhaps this suggests that greater control is necessary in measuring the short dimension. The nonsignificance suggests that the contribution to variance from each effect after the first is negligible. In all cases, once the homogeneity is established, the overall mean and total variance may be used as the best estimate of the population parameters.

Analysis of the second set of measurements is exhibited in tables 8a, 8b, 8c; here again the variation among the set means is less than that among the fragments, and hence, as far as this test is concerned, the deposit is again homogeneous and variation among fragments in a, b, and c axes is the sole source of variation involved. Again, the total variance is the best estimate of the population variance. The variances are not significantly different and hence the results from both sampling designs may be combined. These analyses appear to support the original hypothesis that the size distribution among these fragments is homogeneous and shows no systematic arrangement over the small area sampled.

SHAPE OF THE FRAGMENTS FROM TUSCARORA SCREE

Sphericity of quartz grains and composite fragments of rock (pebbles and boulders) is

usually measured as the ratio between three perpendicular axes (Krumbein, 1941); it is of some interest to determine the sphericity of fragments of scree to estimate the range in sphericity available in the source area. It has been our practice for some time to avoid ratios of axes in microns and to use linear compounds of the various axes in phi units to determine the sphericity relationship (Griffiths, et al., 1955; Griffiths, 1957; Hulbe, 1957). In this case variation in "shape" is summarized as a measure of the relationship between all three pairs of axes in terms of simple linear regressions and of intensity of association by the corresponding correlation coefficient. By definition the "a" (long) axis exceeds the "b" (intermediate) axis and both exceed the "c" (short) axis so that various degrees of cut-off in the frequency distributions may occur. Nevertheless, the axial relationship and associations appear to be interpretable geologically and represent a uniform basis for comparison among grains, pebbles, boulders, and fragments.

The relationship between the axes in logarithmic (phi) units is closely linear and generally homoscedastic so that no violent departures from the normal model should be expected.

In the present example there are 88 fragments of Tuscarora Scree, and three axes were measured on each fragment. The "a" axis will hence forth be designated as X_1, the "b" axis as X_2, and the "c" as X_3. Then, the analysis consists of finding the linear regression relationships among X_1, X_2, X_3 and their degree of association as measured by the relevant correlation coefficients of zero order (r_{ij}). The relationships are displayed as bivariate scatter diagrams in figures 3, 4, and 5 and the relevant statistics are given in table 10.

It seems clear that the "a" (X_1) and "b" (X_2) axes are most closely related and the scatter around this regression line is, as would be expected, least; "a" (X_1) and "c" (X_3) axes are least closely related and the scatter is much larger around this line of regression. The "b" (X_2) and "c" (X_3) relationship is intermediate.

The line $Y=X$, or $Y=\alpha+\beta X$, where $\alpha=0$ and $\beta=1$, is inserted in each diagram and it can be seen that the observed regression lines all fail to pass through the origin (where $\alpha=0$), and β is always less than unity. As a result, the regression line is not parallel to the 1:1 line. This imples that whereas by definition Y could equal X but not exceed it, in actual fact Y is less than X in all cases and tends to become a smaller fraction of X as X increases.

COMPARISON WITH THE RESULTS FROM OTHER DEPOSITS

A number of results using similar analytical procedures are summarized in table 11, including pebbles from a gravel (Olean, N. Y.), fragments of scree (Tuscarora), quartz grains from a beach and dune (Hulbe, 1957), and quartz grains from a Pennsylvanian quartzite (Homewood sandstone, Central Pennsylvania).

With one exception (Olean pebbles) these examples (table 11) show that the relationship among a and b axes is closest and gradually decreases through b and c to a and c. This is, of course, what one would expect from the definition of the axes and what we know of the variation in shapes of individual grains and fragments in sediments. However, the range among a and b axes (from 0.80 to 0.95) is not large and the variation in the bc and ac associations is larger. Hulbe was able to make use of these features in differentiat-

TABLE 9.—*Comparisons of means and variances among sampling designs*

	Mean	Variance	n	Variance Ratio	Degrees of Freedom
a axis	−8.706	0.6219	64		n_1/n_2
a axis	−8.667	0.9984	24	1.605	24/64
b axis	−7.897	0.8057	64	1.063	64/24
b axis	−7.959	0.7576	24		
c axis	−6.821	0.6871	64	1.222	64/24
c axis	−6.932	0.5620	24		

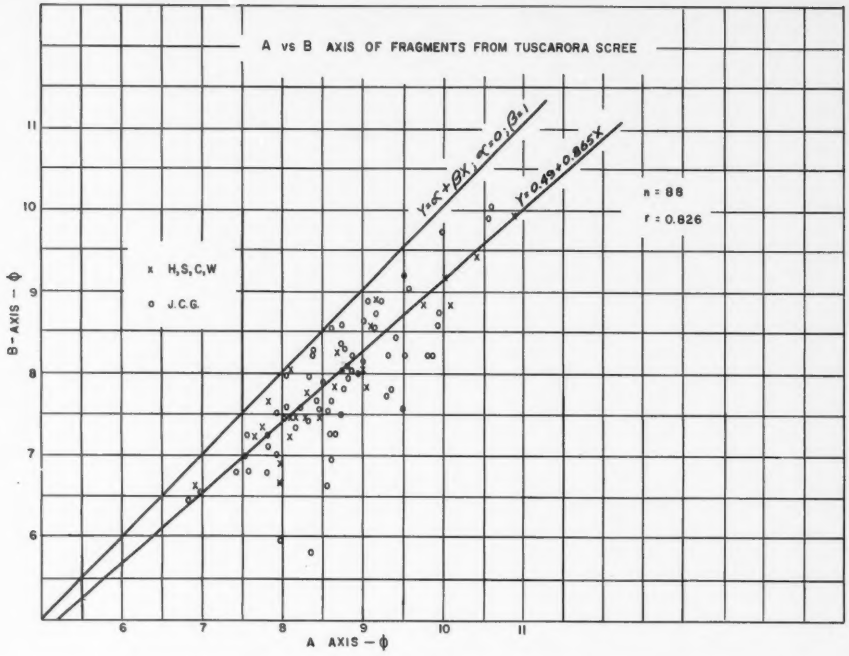

FIG. 3.—Relationships among pairs of axes in fragments from Tuscarora Scree: a versus b axis.

TABLE 10.—*Correlation and regression relationships among the three pairs of axes of fragments from Tuscarora Scree*

Axis Pair	Correlation Coefficient (r_{ij})	Linear Regression $Y=a+bX$
X_1X_2=a vs. b	0.826	$X_2=0.490+0.865X_1$
X_1X_3=a vs. c	0.669	$X_3=1.097+0.656X_1$
X_2X_3=b vs. c	0.689	$X_3=1.838+0.608X_2$

TABLE 11.—*Association among the axes in various sedimentary deposits*

	a vs. b	b vs. c	a vs. c
Olean Conglomerate	0.812	0.570	0.603
Tuscarora Scree	0.826	0.689	0.669
Pedro Beach (Calif.)[1]	0.836	0.773	0.709
Pedro Dune (Calif.)[1]	0.802	0.633	0.628
Homewood Quartzite			
Outcrop 1	0.918	0.900	0.867
Outcrop 2	0.957	0.945	0.923
Outcrop 3	0.947	0.932	0.926

[1] (Hulbe, 1957).

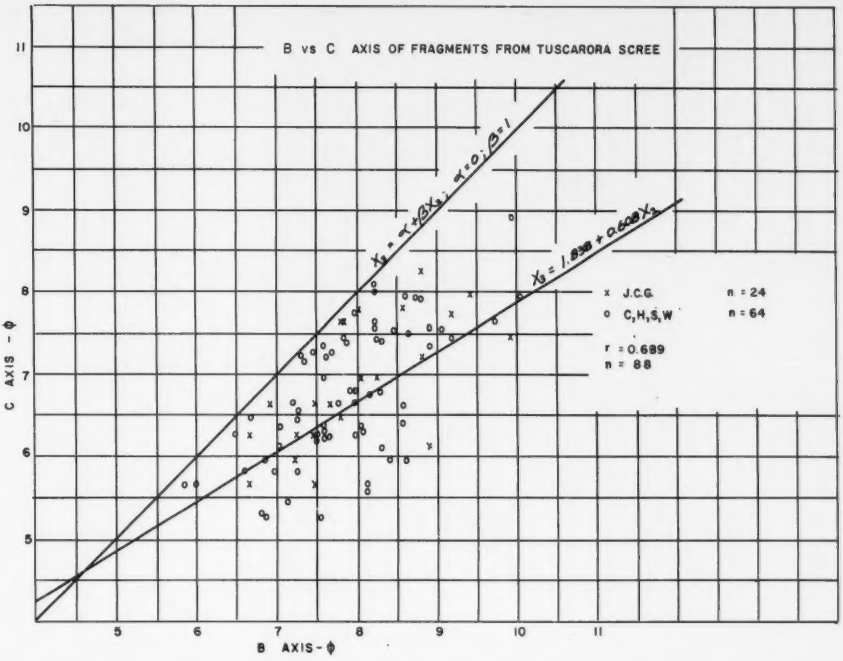

FIG. 4.—Relationships among pairs of axes in fragments from Tuscarora Scree: b versus c axis.

ing quartz grains in a beach sand from those in a neighboring dune (Hulbe, 1957).

Exceptions to the general rule in the above table are also of interest because it seems likely that the shapes of pebbles are characteristic of the agency which deposited them; in this case the variation in shape in the Olean pebbles may be characteristic of a beach environment (i.e., in beach pebbles bc may be less closely related than ac. The quartz grains from the Homewood Quartzite of Central Pennsylvania show much closer relationships among bc and ac than the other sediments, and this may be a reflection of the degree of intensity of selective sorting which they have undergone.

This kind of analysis appears to be somewhat more sensitive to shape changes than previous approaches, and since it can be generalized to three dimensions (or more) it is somewhat more realistic; a set of grains approaching spheres would, of course, yield very high values of r_{ij} for all three axes. A flat, disc-like grain would be characterized by high r_{ab} and much lower r_{bc} and r_{ac}. Thus by combining these three features it is possible to study shape variation and because r_{ij} is dimensionless it is independent of size.

The final step is to compare variation in all three axes simultaneously; this adds more information to the designation of shape and its variability. In a sphere, for example, r_{ab} would become zero if adjusted for variation in c because a, b, and c vary similarly. On the other hand if variation in c were independent of variation in a and b, then the r_{ab} relationship would be unaffected by adjustment for c. The adjustment is performed through the use of the partial correlation coefficient, $r_{ab.c}$ for example, which measures the relationship among a and b axes with the effect of variation in c removed (held constant). The three original and adjusted coefficients for Tuscarora Scree frag-

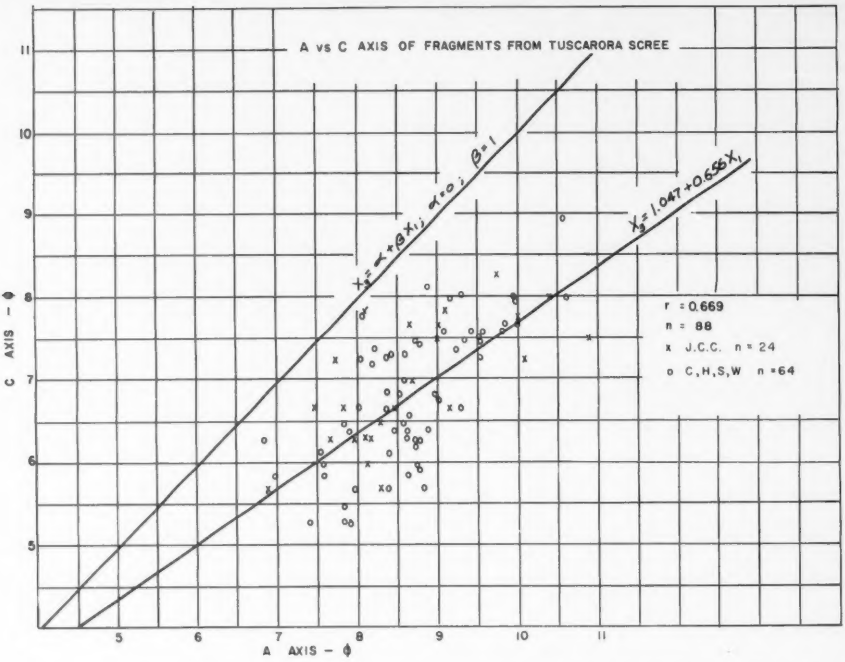

FIG. 5.—Relationships among pairs of axes in fragments from Tuscarora Scree: a versus c axis.

ments are:

$r_{ab}=0.826$	$r_{bc}=0.689$	$r_{ac}=0.669$
$r_{ab\cdot c}=0.678$	$r_{bc\cdot a}=0.777$	$r_{ac\cdot b}=0.245$

The adjustment is performed by means of the definition of a partial correlation coefficient in terms of the zero order coefficients as follows (Fisher, 1948, p. 188):

$$r_{12.3}=(r_{12}-r_{13}r_{23})/\sqrt{(1-r_{13}^2)(1-r_{23}^2)}$$

In this case where c is held constant there is a decrease in the association between a and b; when b is held constant there is an increase in the association between b and c; when b is held constant there is a marked decrease in the association between a and c. It is evident that all three axes are inter-related even in fragments from a scree and it may be hazarded that subsequent history of such materials leads, in general, to intensifying of these inter-relationships. Small variations in this intensification are very senstive reflections of the interplay of selective sorting on the range of variation in shape (axial associations) and, by inference, reflect changes in the depositional environment.

CONCLUSIONS

On the basis of this experiment it may be concluded that variation in lengths of axes of fragments (i.e., grain size and sorting) in this Tuscarora Scree resembles very closely the variation in pebbles and quartz grains in other detrital sediments.

As far as can be determined there is no systematic arrangement in this part of the scree and the variation pattern is homogeneous. For future experiments, then, the largest source of variation is that among fragments so that the number of measurements should be increased at this level.

The shape of these scree fragments, as represented by the relationships among the axial lengths, is similar in degree and kind to that found in detrital particles of other de-

posits and, for analysis of variation in shape the intensity of the associations among the axes, appears to offer the greatest source of information on change in shape with environment.

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FIGS. 1–2, SEPTEMBER, 1959

DETECTION OF LOGNORMAL SIZE DISTRIBUTIONS IN CLASTIC SEDIMENTS[1]

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ABSTRACT

The determination of the form of grain size distribution functions in clastic sedimentary rocks involves a certain amount of subjective interpretation. The subjectivity can be minimized, though not eliminated, by various analytical and interpretive methods. If the object of studying a size distribution is the interpretation of sedimentary history, probably the best method consists of making approximate analyses of a large number of samples and comparing theoretical distributions with whatever data are capable of being tested.

INTRODUCTION

During the past several years a continuing study has been conducted at The Rice Institute concerning the interpretation of grain size distributions of clastic sedimentary rocks and some of the individual heavy minerals contained in them. In the course of the work, a number of techniques for measuring and interpreting data have been developed and used. Many of the descriptive and interpretive techniques are well known, but some are not widely used and may be unfamiliar to other investigators. Furthermore, the use of any method of measurement or interpretation involves the acceptance of certain broad assumptions or basic philosophic approaches. Consequently, the present paper is written for the purpose of summarizing the methods currently used for the study of size distributions and discussing the implications of some of these methods.

The determination of a mathematical function which describes grain size distributions in clastic sedimentary rocks is of importance for two reasons. First, if the form of a distribution function is known, then the processes which caused or were connected with its generation can be hypothesized. For example, if the commonly found lognormal distribution can be interpreted as resulting from the action of a random combination of processes (e.g., see discussion in Aitchison and Brown, 1957), then the presence of such a distribution in a sediment might indicate the absence of non-random combinations of geologic forces (such as dominance of elutriating currents) during transportation and deposition. Such a conclusion is neither necessarily nor unequivocably true, but it does provide a working hypothesis for the interpretation of sedimentation history. A second reason for determining a distribution function is that the quantitative description of geologic information is of value whether or not its "significance," a difficult word to define, can be understood at the present time.

The lognormal distribution is commonly used to describe sedimentary size distributions because it is a good approximation to many size frequency curves. The significance of the distribution and departures from it has recently been discussed by Harris (1958a, 1958b), Mason and Folk (1958), and Rogers (1958). Tanner (1958) discusses the significance of distributions related to the lognormal. Miller and Goldberg (1955) have discussed the implications and the detection of lognormality in geochemistry.

The object of the present paper is to investigate the methods used to detect lognormality of sedimentary grain size distributions and to discuss the implications of the different techniques.

METHODS OF MEASUREMENT

Size distributions of clastic sediments can be measured in a variety of ways, including sieving, grain counting, and settling techniques. Settling methods provide only indirect measurement of grain size; therefore the present discussion is concerned solely with sieving and grain counting. Prior to

[1] Manuscript received January 16, 1959.

describing these techniques, however, it is necessary to establish exactly what is meant by the terms "grain" and "size."

A "grain" in a clastic sediment is presumably a particle which is deposited as such from a transporting medium. Thus, in a sandstone, particles of quartz, feldspar, and rock fragments, have probably been transported independently of each other, and each individual particle is classified as a grain. In a sandstone, measurement of size distribution is made on the entire material after mechanical disintegration of the rock to these individual fragments.

In rocks other than sandstones, however, the recognition of grains may be quite difficult. The individual clay particles in a clay shale probably settled from suspension in aggregates rather than as individual grains (they have also probably undergone extensive diagenesis). Consequently, disaggregation of a shale to individual clay particles and size analysis of these grains may provide little or no information concerning sedimentation history. The sand- and silt-sized quartz and other mineral grains scattered throughout many shales certainly are not in hydraulic equilibrium with the individual clay grains, but they may have the same range of settling velocities as the clay aggregates which have accumulated to form the apparently homogeneous clay fraction of the shale. Another problem in disaggregation of samples is met with in studying weathered products and partly disintegrated material. Most such rocks require additional disintegration in the laboratory before they can be passed through sieves, but it is never certain just how vigorous this disaggregation can be without destroying the "grains" whose distribution is to be measured.

The term "size" is equally as difficult to define as the term "grain." True size should mean volume, but volume can be determined directly only for very large fragments whose displacement in water or some other fluid can be measured. Both sieving and grain counting involve linear measurements, which can be considered only as approximations to grain volume. Whereas grain counts generally involve the long diameter of a grain, sieving measures a dimension close to the intermediate diameter. Probably the best definition of "size" states that it is a grain parameter, determined by any method used to measure it. In this sense, size is probably little, if at all, different from other physical parameters.

Size distributions of most sand-sized sediments are determined by sieving. The method is rapid, generally reproducible, and meaures the distribution of an enormous number of grains. Sieving of bulk material determines the size distribution of the "light" minerals of a sediment because these minerals comprise by far the largest portion of most sediments. Only two precautions need be mentioned in connection with sieve analysis. One is that large samples give incorrect results because the sieves may become choked by sediment, preventing some grains of smaller size than the sieve openings from passing through. Samples should not be made too small, however, for they will then neither provide sufficient material on individual sieves for accurate measurement, nor will they be as representative of the whole sediment as is a larger sample. Generally, samples of 30 to 50 grams can be sieved satisfactorily, and for coarse sediments, 100 to 200 grams can be handled. A second precaution in sieving sediments concerns the interpretation of analyses in which the material recovered does not approximate 100 percent of the original sample. In sieving sands, especially unconsolidated ones, the problem generally does not arise, for in such cases the recovery is close to 99.9+ per cent. In some sediments, however, particularly those containing clay or organic matter, some material may be lost either by escaping as dust during some part of the preparation or sieving process, or by sticking to screens during sieving. In calculating size distributions for such sediments, it is necessary either to make some assumption concerning the size of the material lost, or to consider the recovered material as 100 percent of the sediment. Either method is certainly satisfactory in determining median sizes, sorting coefficients, and other properties in routine analysis of large numbers of sediments. But in attempting to determine the exact nature of the size frequency distribution, probably the most rigorous approach would be to ignore data from sieve analyses with less than effectively

100 percent recovery and to consider the information not usable.

Measurement of size distributions by grain counting provides a direct measurement of numerical frequency distribution. It has been repeatedly shown that a simple relationship exists between lognormal distributions by weight (as measured by sieve analysis) and by number, and mathematical proof is given by Aitchison and Brown (1957). A lognormal distribution by weight percentage is also lognormal by number percentage and has the same standard deviation (sorting coefficient) as the numerical distribution; a weight distribution has a "median size" coarser than that of the corresponding numerical distribution. Except in geologic sciences, the term "frequency distribution" almost invariably refers to numerical distributions (e.g., in studies of grain sizes in photographic emulsions, colloids, etc.).

Grain counts are commonly made on approximately 300 grains mounted in balsam. The number of grains measured is based on recommendations by Dryden (1931). In order to provide extreme accuracy (in the range of 0.1 percent), so many grains must be counted that the results are generally not considered worth the effort. Rather than count exactly 300 grains, it is probably more accurate to make traverses at set intervals which will sample an entire slide and cross approximately 300 grains. Most slides are reasonably homogeneous with regard to scattering of different sized grains; and, slides which are not homogeneous can be readily recognized. Consequently, sufficient accuracy can probably be obtained by counting an even 300 grains. One possible modification is that if the 300th grain is in a small cluster of grains of different sizes, it would seem more accurate to count either all of the grains in the group or none. Thus, the count would range within a few grains of 300. The words "sufficient accuracy" are used above because it is not possible to establish absolute identity of a frequency distribution with the lognormal or any other distribution function; only a close approximation can be determined. The data from counts of 300 grains should be recorded only to the nearest one-third percent.

PLOTTING DATA

Data obtained from the various measurements must be compared with theoretical lognormality and tested for deviation from it. One way in which to accomplish this is to plot the data on logarithmic probability paper, on which cumulative lognormal curves plot as straight lines. Deviations from straight lines are far more readily detected than deviations from other types of curves. Size data may be converted to phi units (Krumbien, 1938) and plotted on arithmetic probability paper. This method is of merit in providing arithmetically distributed data for subsequent statistical analysis.

One question which must be posed concerns the use of the lognormal, rather than some other, type of normal curve. Any unimodal distribution can be converted to a normal curve by appropriate selection of a size scale, although some very peculiar scales may have to be chosen in order to fit some data. The reason for using log size is that this scale commonly provides the closest approach to normality. Probably a more fundamental significance can be attached to the use of the logarithm, although it is not certain what that significance might be.

Once the data have been plotted and a theoretical curve drawn for comparison, various statistical tests may be used to determine significance of deviations from the theoretical curve. The familiar chi-square test is commonly used, though others, possibly more suitable, are described by Aitchison and Brown (1957). These tests cannot establish with certainty that any distribution is lognormal. Even for data which fit lognormality precisely (chi-square for deviation equals 0), the only conclusion which can be drawn is that there is no evidence that the distribution is not lognormal. It is always possible that a set of data which appear to be lognormal may fit some other distribution better, though if chi-square for deviation from lognormality is 0, no better fit can be found. The subjectivity which enters the final decision as to whether certain data are lognormally distributed or not is unfortunate but inescapable.

The above discussion of testing for deviations is based on the assumption that a theoretical lognormal curve can be drawn

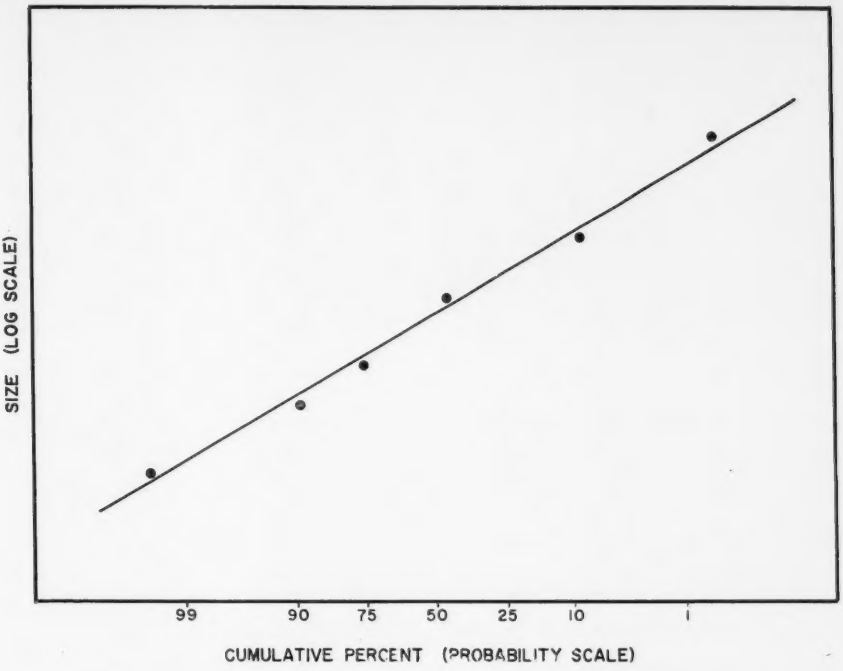

FIG. 1.—Size distribution data compared with theoretical lognormal curve.

as a best approximation to any given data, and the data can be compared with it. Construction of this theoretical curve, however, is not a routine problem. Consider, for example, the two sets of data illustrated in figures 1 and 2. In figure 1, the plotted points are not all colinear, and chi-square or other tests might indicate "significant" deviation from the theoretical lognormal curve which has been drawn for comparison. Nevertheless, in figure 1 it seems apparent that the theoretical comparison curve must be drawn exactly as it is, with the median and slope as shown. In figure 2, however, a different situation exists. No difficulty would be experienced in drawing a theoretical curve (curve A) which would approximate most of the points very closely and would yield a very low chi-square value for deviation of the actual data. The difficulty in figure 2 is that geologic deviations from lognormality appear to occur far more commonly in the tails of distribution curves than in their central portions. Consequently, the data might actually not be lognormal even though statistical tests would indicate lognormality simply because the very small percentage of material in the tails contributes a comparatively small amount to chi-square or other statistical values. In figure 2, it might be more accurate to compare the data with theoretical curve B than with curve A, even though the statistical deviations from curve B are greater than those from curve A. Statistical methods for drawing curves are of little value here because a subjective judgement has been, and must be, made concerning the importance of different parts of the data. Logically, if one is trying to test for lognormality and the data plot as in figure 2, it would be most accurate to forget the data and simply state

that no decision can be made concerning them. Partly for this reason it is almost always necessary to measure the size distributions of a number of samples from a sediment before drawing definite conclusions.

DESIRED ACCURACY

Considerable effort is involved in determining the size distribution of a clastic sediment, and it is always necessary to decide just how much work can profitably be done in order to reach the desired goal. The answer, of course, depends on the purpose for which the work is done.

If the object of investigating the size distribution of a sediment is to determine the form of a distribution function, then a great deal of work should be expended on collection of the numerical data. In determining, for example, the size distribution of halide grains in photographic emulsions, some workers have counted close to 100,000 grains. The purpose in such work is a practical one: either the researcher wishes to determine the response of the emulsions to light, or he wishes to test a very practical explanatory model concerning the nucleation and growth of crystals in a homogeneous medium. Even 100,000 grains do not provide 100 percent certainty, but the sample size is effectively infinite, and if the data fit a lognormal distribution, it can be assumed that the distribution is probably valid.

If the object of studying size distributions in a sediment, however, is to use the distribution to interpret the history of the sediment, then it is questionable whether there is any value in counting 100,000 grains, or even 1000. The model used by Kottler (1950) to explain the lognormal distribution of grains grown in a homogeneous medium is far more precise and rigorous than any model which can currently be proposed to

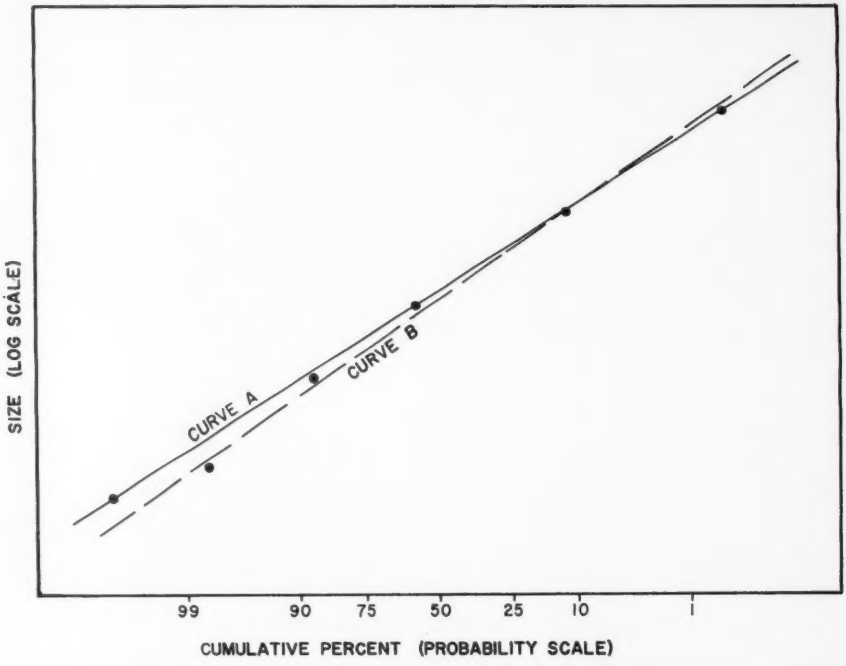

FIG. 2.—Size distribution data compared with two possible theoretical lognormal curves.

explain natural sedimentary size distributions. Precise interpretation of sedimentary history is hindered by the large number of variables involved in the erosion, transportation, deposition, and diagenesis, of sedimentary material. For example, the deficiency of fine material (relative to log-normal) in some current-washed sediments may be attributed to elutriation, but it might also be explained by a deficiency in supply of material of that size, or by many other reasons. Possibly a limited, though accurate, effort on a large number of sediment samples is more productive than intense effort on a few samples if the object of the work is the interpretation of geologic history. It seems probable that the effort to achieve extreme accuracy in some geologic investigations is unneccessarily time-consuming. The question as to the amount of effort which can profitably be expended must be answered by the individual researcher in terms of his own interests and his opportunity for pursuing them.

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FIGS. 1–4, SEPTEMBER, 1959

SAMPLE COMPONENTS OBTAINED BY THE METHOD OF DIFFERENCES[1]

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ABSTRACT

The method of first and second differences permits, in certain instances, the detachment of major components from zig-zag curves. Because of the overlap between components, and other factors, the method is commonly subjective in part and therefore not rigorous. Three applications are shown: two dealing with modern shoreline marsh samples and one dealing with an Eocene sand. Recombination of the detached components provides curves which, generally, deviate from the original less than two percent. After components have been obtained, it is still not possible to identify the site of deposition of the sample. Additional applications are known from geomorphology, stratigraphy, river hydrology, and other branches of earth science. Because the results in geomorphic problems can be verified in the field, it is thought that the analysis of sediment samples is also meaningful.

On probability paper many sets of sedimentary data plot as zig-zag curves rather than as straight lines. It has been suggested (Tanner, 1958) that zig-zag curves may be composed of two or more Gaussian components, and it is believed (Tanner and Bagg, in preparation) that the method of differences may permit an investigator to determine the nature and properties of the components.

The method of differences is similar to the method of derivatives which permits one to obtain approximations to means (maxima on the smooth-curve histogram), percentages (minima), and standard deviations (determined from inflection points). First and second derivatives are difficult to get, however, even if the equation of the curve can be written, which is commonly not practicable. First and second differences are relatively simple to obtain from the array and yield information almost as useful as that determined by derivatives.

First differences are calculated by subtracting the percentages in the array from one another in orderly fashion as illustrated in table 1. Second differences are then computed by manipulating, in a similar fashion, first differences. The results are plotted in figure 1. Data for this illustration were taken from a marsh sample examined by Waskom (1958).

Figure 1 contains the cumulative percent scale, also obtained from the array, on which all values must be read. Significant points are those where first differences (solid line) equal zero. Where first differences equal zero and second differences are negative, means (approximately) appear; where first differences are zero and second differences are positive, proportions or percentages (approximately) appear. An examination of the figure shows that the data seem to contain three components with three means which fall on the original cumulative frequency curve at about four percent, about 60 percent, and about 88 percent.

Percentages appear twice, dividing the array at about seven percent and about 87 percent. From these two divisions one calculates three proportions of about 7,80, and 13 percent respectively. These are shown in figure 1.

Standard deviations may be obtained by noting where the means, and adjacent points where second differences are zero, fall on the cumulated frequency curve. These values, read finally in phi units, give a standard deviation of between .3 and .45 for the central component (80 percent). Apparently the disproportionate ratios between components, plus the overlap, prevent reasonable standard deviations from being read for the two lesser components.

With three means, three percentages, and a standard deviation (more or less), one can plot the main component (80 percent). The differences between this main component

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TABLE 1.—*Cumulated percentages, percentages, and first and second differences for the modern marsh sample, 57WB7. Differences, as computed here, are plotted in figure 1. The complete analysis is shown in figure 2.*

Cum.%	%	I	II
.12	.12		
		.56	
.80	.68		2.20
		2.76	
4.24	3.44		−4.20
		−1.44	
6.24	2.00		1.37
		−.07	
8.17	1.93		1.67
		1.60	
11.70	3.53		2.52
		4.12	
19.35	7.65		−.17
		3.95	
30.95	11.60		3.60
		7.55	
50.10	19.15		−6.43
		1.12	
70.37	20.27		−8.26
		−7.14	
83.50	13.13		−3.50
		−10.64	
85.99	2.49		9.39
		−1.25	
87.23	1.24		1.32
		.07	
88.54	1.31		−.25
		−.18	
89.67	1.13		.15
		−.03	
90.77	1.10		−.21
		−.24	
91.63	.86		.96
		.72	
93.21	1.58		4.16
		4.88	
99.70	6.49		

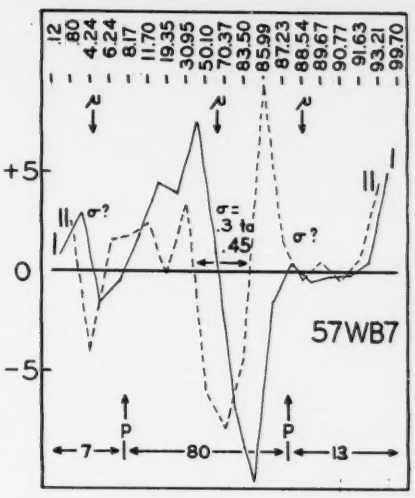

FIG. 1.—First difference (solid line) and second differences (dashed line) for a modern marsh sample. The array of data is given in table 1. Critical points occur where first differences are zero; at those points, negative second differences indicate means (μ), and positive second differences, percentages or proportions (P). Both parameters are read against the cumulated percentage scale at the top: (a) percentages, to be obtained by subtraction, as shown at the bottom, and (b) means, to be located on the probability plot (fig. 2). Standard deviations are read, in certain favorable instances, from the same critical points, taking the distance to the nearest zero second difference (see arrows marked σ); the values, however, must be read in phi units off of the probability plot.

and the total curve, obtained by subtraction and weighting with the percentages given, determine the other two components. By this manipulation, three Gaussian distributions are defined. Unfortunately (as is commonly the case), they fail to recombine properly, and therefore must be adjusted slightly. The adjustment, designed to eliminate as much as possible the obvious errors provides a final analysis as given in table 2.

Figure 2 shows the three adjusted components, the original data (solid zig-zag curve), and the combination of the three components (dashed zig-zag curve). The departure of the computed curve from the observed curve is generally less than one percent, and at no place is it more than about three percent. This departure could be reduced even more by splitting the main component into two lesser ones, but the present writer feels that the three-component analysis is adequate to a computational accuracy of about three percent.

A second example, also a marsh sample taken by Waskom (1958), is shown in figure 3. Again, three components are detached, with the proportions shown. The departure is at no place as much as two percent. A third example from the (Eocene) Nanafalia Formation outcrop near Andersonville, Georgia, is given in figure 4. Four components are indicated, including one which

TABLE 2.—*Analysis of sediment sample data, marsh sample, No. 57WB7.*

Per Cent	Mean		$D_k \approx \sigma$	So
	Phi	Mm.		
9	0.75	0.6	0.35	1.17
77	2.20	0.22	0.35	1.17
14	4.25	0.05	0.65	1.38
100	2.23	0.21	0.58	1.32

consists of clay pellets or flakes. The departure nowhere exceeds three percent, and assumption of a fifth component could reduce even that figure.

Similar analyses, carried out on geomorphic data, provide distributions which are identifiable as such in the field: for instance, erosion surfaces, lithologic differences, and various pediment levels. For circular data, such as cross-bedding directions (Tanner, 1959), this type of operation commonly permits computational rigor in separating modes which appear to have paleogeographic significance. Calculations on river flow data, checked by unit hydrograph analysis, also show that detachment of modes can be meaningful.

For the three examples given here, it is more difficult to find a positive meaning. The clay pellet distribution in figure 4 can be verified by visual inspection; the other modes probably cannot. In each of the three

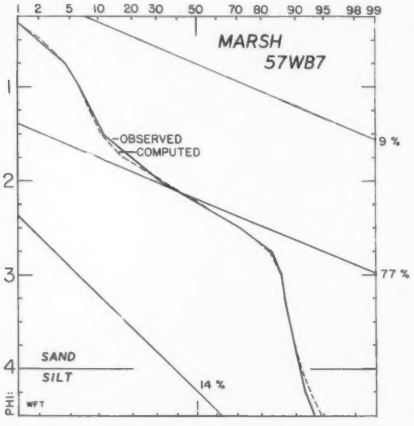

FIG. 2.—Probability plot for the marsh sample treated in fig. 1 and table 1. The three chief components are given as straight solid lines, contributing 14%, 77% and 9% to the total. The dashed line is the combination of the three.

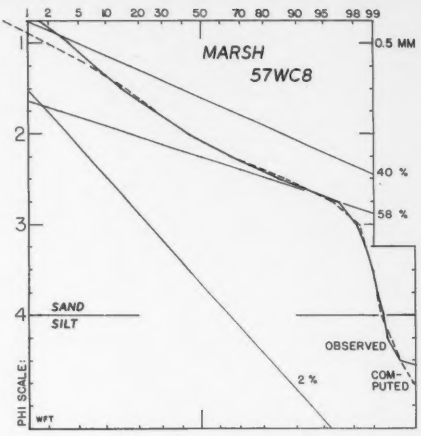

FIG. 3.—A second marsh sample, showing three components contributing 2%, 58% and 40% to the total. Components with this much overlap may be difficult to separate.

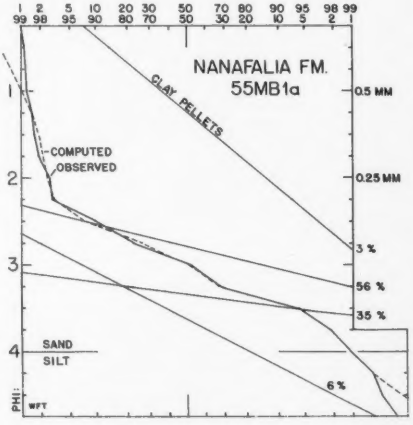

FIG. 4.—Probability plot for a sand sample from the Nanafalia formation, near Andersonville, Georgia. Four components are shown, contributing 6%, 35%, 56% and 3%. The latter is chiefly clay pellets. The dashed line is the combination of the four.

cases three components have been separated by analysis, two of which are quartz sand, and one of which is quartz fine-sand-and-silt. The overlap is great enough to prevent separation by screening. Whether some other technique might achieve this aim is not now known. The several components are to be accounted for by ordinary processes. Similar concepts have been put forward recently by Groot (1955) and Folk (Folk and Ward, 1957, p. 25; Mason and Folk, 1958, p. 223).

The components indicated in the figures are not to be confused with the major components (gravel, sand bed load, sand-and-silt wash load) named in a previous paper (Tanner, 1958), although it is possible that the C-point, shown in that paper, also appears in the present figures.

Unfortunately, this type of analysis has not permitted identification of site of deposition. Zig-zag curves having three or four components have been obtained from river, dune, beach ridge, marsh, lagoon, protected swash, open swash, and offshore sands.

Several words of caution, based on experience in analysis of many zig-zag curves representing various types of earth science data, should be stated here. For various reasons the procedure does not always work. There is no guarantee that the solution, when obtained, is unique; although the evidence suggests that it may be. The operation is, in part, subjective; its success is related, to some extent, to the skill of the analyst. And of course, the basic assumption—that practically all components have Gaussian (or, alternatively, log-normal) distributions—is certainly not known to be true.

However, the procedure is effective in many cases; and the results are reasonable enough and interesting enough to make the operation worth while.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 29, NO. 3, PP. 412–424
FIGS. 1–9, SEPTEMBER, 1959

THE MECHANICAL COMPOSITION OF SOME INTERTIDAL SANDS[1]

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ABSTRACT

The effect of wave action without longshore drift on the mechanical composition of sands is shown to consist chiefly of the removal of the finest grades from the material on the beach face. The mechanical composition of the parent material, not the sorting process, is the main influence controlling the grading range of the coarser fractions. The fields for plots of median versus logarithmic equivalent grade for dunes and beach deposits are different, and together with plots of the ranges of grades coarser and finer than the logarithmic equivalent grade, serve to readily distinguish between the two environments. In a re-examination of the shore zones of McKelvey, evidence is given for splitting his zone D into a beach face (DA) and beach flat (DB).

INTRODUCTION

When Udden extended his studies of the mechanical composition of sediments from dune sands to other deposits, the close similarity between the mechanical composition of intertidal and dune deposits became apparent (Udden, 1914). Although it is easy enough to separate glacial till from dune sand, the separation of intertidal and dune deposits has remained a problem.

Recent studies of dune sand from the Suez Canal Zone (Harris, 1957; 1958B) have shown that the mechanical composition within a given microenvironment varies little. New methods have permitted the separation of the characteristic mechanical composition of each microenvironment. The differences between one microenvironment and another are considerable and this may be the explanation of the considerable variation found by earlier workers (e.g. Udden, 1898; Bagnold, 1941).

The work of Mason & Folk (1958) at Mustang Island, Texas, suggests that dune sands and beach sands have a different mechanical composition at any one place. This study produced results similar to those of Harris (1957) regarding wind erosion, though their methods of study of these results were different. Krumbein (1938) demonstrated that beach sands vary little over the beach face; that is, single samples can be expected to give a good idea of the composition of the face of the beach. Maximum variability is found in the direction at right angles to the contours. The question arises as to whether beach deposits can be confused with dune sands in studies of mechanical analyses. If so, mechanical analyses cannot be used on their own for the recognition of aeolian microenvironments. In this paper, an attempt will be made to apply the new methods of study to actual examples of intertidal sands.

[1] Manuscript received March 13, 1959.

BEACH ZONES

In 1940, McKelvey published a study of the beach deposits of the shores of Trout Lake, Wisconsin (McKelvey, 1940). He recognised eight zones based on field and laboratory work, representing most of the zones possible on a shore. Only rarely are all these zones found in one beach profile.

In view of the exceptionally detailed study of the subaerial and intertidal zones, the present writer has attempted to define the detailed changes from McKelvey's histograms in table 1. These zones will be used for descriptive purposes in the ensuing studies.

BEACH AND DUNE SANDS IN A SHORE SECTION

In order to check the relationship between beach and dune sands in a single section, a study was made of the shore at Bir Themada (G.R. 7465, 7743, Egypt Red Grid) on the west coast of the Gulf of Suez.

The beach section is shown in figure 1, together with the sampling sites. It is a beach of accretion, with dune sand trapped by the ground moisture at the back of the beach feature, producing a line of low dunes. The predominant winds were from the north and south-west and also from the east (be-

TABLE 1.—*The eight shore zones (after McKelvey, 1940)*

Zone Symbol	Zone	Definition	Characteristics (seawards)
A	Cliff	—	Parent material
B	Backshore	Between the high tide mark and the cliff	Reduction in the coarsest and finest grades
C	Swashline	Normal high tide mark	Build up in coarse grades—often a pebble concentrate
D	Foreshore	Swashline to plungeline	Coarsest and finest grades completely removed
E	Plungeline	Normal low water mark	Coarse material reappears
F	Upper cobble Zone	Probably place where waves break	Increase in finer textured material, but no change in coarse grades, therefore ill-sorted
G	Sand Gravel Zone		Increasing but variable deposition of coarsest grades
H	Lower Cobble Zone		Progressively more ill-sorted material. The resulting bimodal effect may be more pronounced than in the parent material

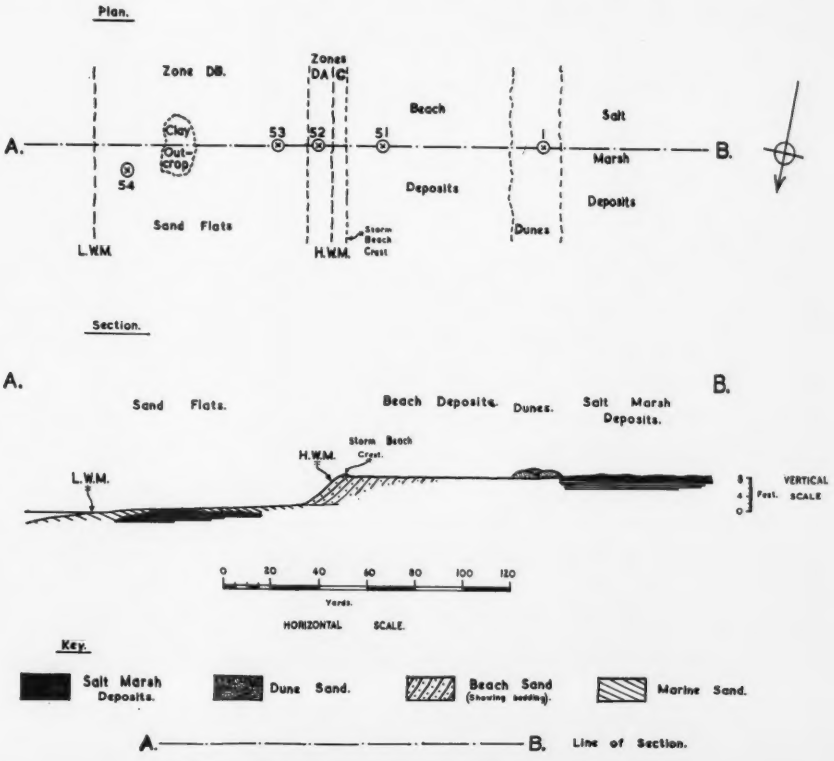

FIG. 1.—The Bir Themada shore profile, Suez Canal Zone.

tween mid-day and 4 p.m.), the last two about balancing each other in effect at this place.

Zone C was only extremely poorly represented and therefore not investigated, while zones A and B were absent. Zone D was divisible into two parts, *viz:* a sloping beach face (zone DA) and a sandy flat (zone DB). Laminations occurred in the beach face of zone DA as in figure 1.

The results of the mechanical analyses of the samples are given in table 2 and figure 2. The parent material was a mixture of wind-blown sand and a little wadi gravel (brought down during periods of flood by the few open wadi distributaries). The characteristics of the curves and associated mineralogy can be summarised as follows:

(1). *Samples 53 and 54.*—The coarser and finer extramodal curves are extremely small, the modal part of the curve making up almost all of it. Mineralogically, it is entirely composed of clear quartz.

(2). *Samples 51 and 52.*—The modal part of the curve is still marked, while the finer extramodal curve remains extremely small. However, the coarser extramodal curve extends over a much wider range of grades, even if the percentage forming it is still minute. Mineralogically, a few small angular flints are present as well as broken shell remains, in addition to the clear quartz grains; these last two forming the bulk of the deposit, while the flints and shell remains form the coarser extramodal curve.

(3). *Sample 1.*—The modal part of the curve is still marked, while the coarser extramodal curve is insignificant. On the other hand, the finer extramodal curve is very marked, as in true dune deposits; it is totally of clear quartz.

The results of plotting the cumulative curves for samples 1 and 51 to 54 on logarithmic probability paper will be found in table 3 and figure 3. As in the aeolian deposits considered above, a series of lines can be seen, showing that these deposits, too, follow a logarithmic probability function. Since the mechanism is similar in both the fluid media, similar lines would be expected to occur. These samples can be divided into three groups:—

(1) *Samples 53 and 54.*—Each of these is represented by three lines, corresponding to the first three lines (A, B and C) found in aeolian deposits. No material appears to be moved by surface creep in these samples. The percentage corresponding to the third line (*i.e.* M and Q—forming the bulk of the deposit) increases towards the low water

TABLE 2.—*Results of mechanical analyses of samples from the Bir Themada, Sandy Cove and Kallia Shore sections*

% passing B.S. sieve no.	Sample No.									
	1	51	52	53	54	2	55	56	57	58
3″	—	—	—	—	—	—	—	—	100	—
1½″	—	—	—	—	—	—	—	—	95.7	100
¾″	—	—	—	—	—	—	—	—	74.3	90.6
⅜″	—	—	—	—	—	—	—	—	52.2	79.6
3/16″	—	100	100	—	—	—	—	S	37.8	61.3
7	—	99.9	99.6	—	—	—	—	h e l l	25.9	28.0
14	—	99.3	98.5	—	—	100	—	l	19.5	6.5
25	100	96.5	93.5	100	—	98.7	100	s	16.8	1.4
40	98.3	88.5	73.9	98.5	100	88.4	99.3	100	15.2	trace
52	60.0	37.5	24.4	59.1	82.1	59.0	69.2	97.5	13.7	—
60	n.d.	13.1	n.d.	30.2	22.6	n.d.	n.d.	n.d.	n.d.	—
80	6.1	n.d.	n.d.	n.d.	n.d.	19.1	6.5	57.8	11.4	—
100	4.4	0.1	0.1	0.5	0.2	9.9	1.7	13.6	10.1	—
200	1.4	0.0	0.0	0.0	0.0	0.1	0.1	0.1	5.5	—
ϕ median	1.70	1.52	1.40	1.78	1.80	1.75	1.85	2.45	—	—
Log.E.G.	1.83	1.58	1.47	1.83	1.84	1.88	1.905	2.425	—	—

For sizes of sieve apertures, see Harris (1957).

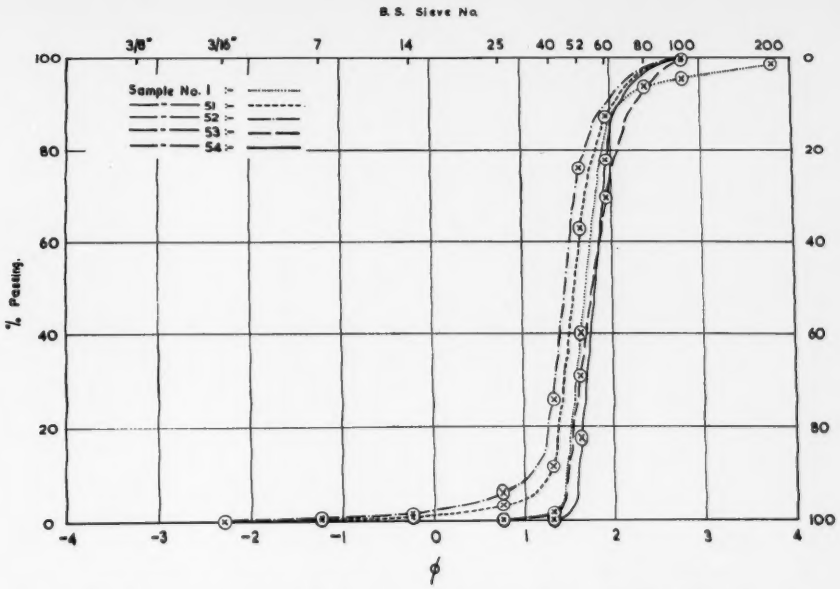

FIG. 2.—Cumulative curves for five samples from the shore profile, Bir Themada.

mark (L.W.M.). There is also a decrease in maximum grain size seawards, resulting in a higher value of ρ for this line in that direction. The ϕ value for the commencement of the material of interstitial type remains fairly constant. In contrast, for interstitial material and material moved in suspension by only the strongest waves, we find that the quantity alters, although range of grain size remains constant. This results in lower values for ρ for these lines towards the sea.

(2) *Samples 51 and 52.*—All the four theoretical lines are present. The sample on the active beach face has a slightly coarser maximum grain size than samples 53 and 54. Once again, the line corresponding to the bulk of the deposit is extremely well graded—about the same as sample 54, but much more so than sample 53 (from just below the beach face). Little variation in actual grading range of the lines in the two samples occurs, and there is no major change in comparison with samples 53 and 54.

(3) *Sample 1.*—This also includes a much greater range of finer grades as interstitial material, while the minimum grain size corresponds to that for normal aeolian deposits. As a result, only data for the two coarsest lines are available. These show that the bulk of the deposit is as well graded as in samples 52 and 54, but the degree of sorting of the fourth line (that due to material moved by surface creep) is much higher than in the samples from the beach face. This is due to a higher percentage being present, yet occupying a smaller range of grades.

Thus there is a marked difference between sample 1 and the others. In regard to these, material moved by surface creep only occurs on the beach face, and not on the beach flat, in this case. This is probably due to the low current velocities except where the waves are actually breaking. The reduction in minimum grain size was noted, while the degree of sorting of the bulk of the deposit decreases between the beach face and near the low water mark (probably due to the shorter period for which it is subjected to wave action).

This difference is further emphasized by

TABLE 3.—*Summary of the results of plotting the samples from the beach sections at Bir Themada and Sandy Cove, Suez Canal Zone on logarithmic probability paper*

Sample	Theoretical Line										
	W					X					
	Line No.	Total Grading (ϕ)	Grading Range (ϕ)	%	$\rho = \%/\phi$	Line No.	Total Grading (ϕ)	Grading Range (ϕ)	%	$\rho = \%/\phi$
1	—	—	—	—	—	X	—	? to 1.90	—	—
51	—	0.10	2.80 to 2.70	1.0	10.0	—	0.88	2.70 to 1.82	14.0	15.9
52	E	0.35	2.85 to 2.50	1.1	3.1	F	0.85	2.50 to 1.65	19.9	23.4
53	—	—	—	—	—	K	0.23	2.95 to 2.72	1.1	4.8
54	—	—	—	—	—	O	0.25	2.95 to 2.76	0.8	3.2
2	W	—	3.80 to ?	—	—	X	—	? to 1.48	—	—
55	K	0.35	3.80 to 3.45	0.9	2.6	L	0.83	3.45 to 2.62	1.6	1.9
56	F	0.31	3.80 to 3.49	1.5	4.8	G	0.42	3.49 to 3.07	1.0	71.0
	Theoretical Line									
	Y					Z				
1	Y	0.40	1.90 to 1.50	52.0	130.0	Z	0.75	1.50 to 0.75	37.0	49.3
51	—	0.70	1.82 to 1.12	80.0	114.3	—	3.41	1.12 to 2.29	5.0	1.5
52	G	0.45	1.65 to 1.20	67.0	148.8	H	3.49	1.20 to −2.29	12.0	3.4
53	L	0.92	2.72 to 1.80	38.9	42.3	M	1.05	1.80 to 0.75	60.0	57.1
54	P	0.85	2.70 to 1.85	21.2	24.9	Q	0.53	1.85 to 1.32	78.0	147.2
2	Y	0.23	1.48 to 1.25	24.0	104.3	Z	1.48	1.25 to −0.23	7.0	4.7
55	M	1.14	2.62 to 1.48	79.5	69.7	N	0.73	1.48 to 0.75	18.0	24.6
56	H	1.26	3.07 to 1.81	89.5	71.0	J	0.49	1.81 to 1.32	8.0	16.5

examination of the graphs in figure 4. The fields of the marine samples are quite distinctive from that of sample 1. The latter is from the surface of salt marsh deposits, which lack the coarser grades.

These results demonstrate that zone D of McKelvey may be divisible into two parts in certain cases:

(1) Zone DA, characterized by the laminated and wedge-bedded deposit of the actual beach face.

(2) Zone DB, characterized by the shallow intertidal shelf between the beach face and the low water mark, with no wedge-bedding present, but with ripple-marking common.

NATURE OF THE SORTING ACTION OF WAVES

We have noted a variation in degree of sorting from zone DA to zone DB. What is the effect of wave action on the mechanical composition of sand?

A study of this effect was carried out at Sandy Cove (G.R. 7385, 8726, Egypt Red Grid) on the shores of Lake Timsah, 5 miles south of Ismailia, Suez Canal Zone. There is an alteration in level of about 2 feet when ships pass. The latter cause a piling up of waves across the shallow offshore area, about 12 inches higher than the mean (just

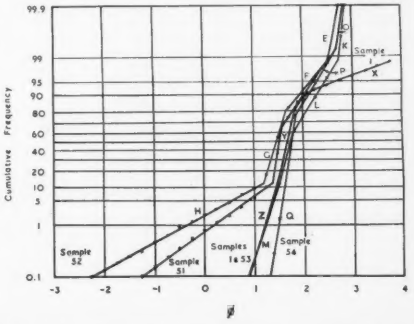

FIG. 3.—Logarithmic probability diagram of the samples from the shore profile, Bir Themada.

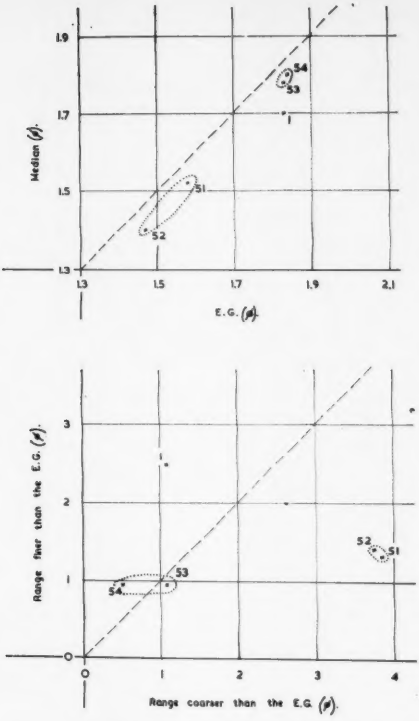

FIG. 4.—Graphs showing the relationship between median and logarithmic equivalent grade and the grain size range coarser and finer than the logarithmic equivalent grade for the samples from the shore profile, Bir Themada.

below the storm notch) while between each ship passing, the waves dropped to about 12 inches below the mean (this being the level of the surface of the water when no ships were passing). It is non-tidal, and the waves were hitting the shore so that their direction of travel was at right angles to the beach at the time of sampling. No longshore drift was occurring. A detailed section of the beach is shown in figure 5. The shore is one of erosion of a low terrace of wind-blown sand (approximately 4 feet above the mean lake level) and the present sand-dunes.

Note that the ripples cease abruptly at notch "A," in a groove parallel to the erosion surface. It was observed that the waves broke exactly over this groove, disturbing all the surface material above it, but not below it. "A" moved according to the size of the waves, ripples forming immediately behind it as wave size decreased. Its normal position was at the foot of the beach slope, where the waves usually break at both mean and high levels, but at low levels it usually migrated down the marine flat a short way.

Table 2 and figure 6 show the results of mechanical analyses of the samples taken from this profile.

Sample 2 has already been discussed with the other dune sands. It is a fairly typical aeolian sand.

Samples 55 and 56 are slightly better graded than sample 2. Thin-shelled Lamellibranchs and some thick-shelled forms are

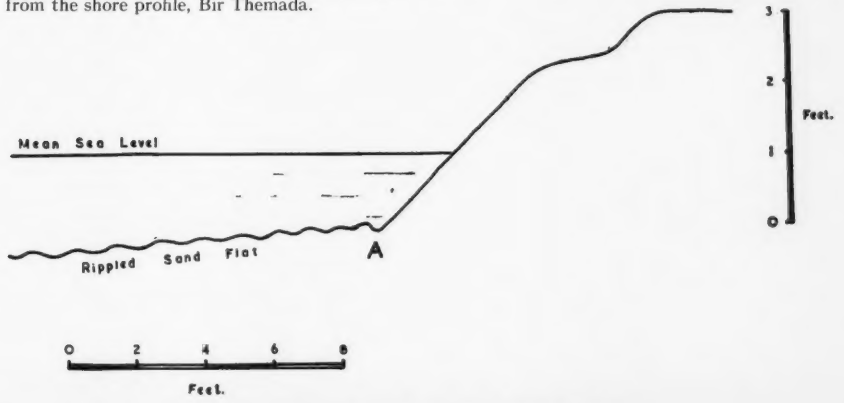

FIG. 5.—Shore profile at Sandy Cove, Suez Canal Zone.

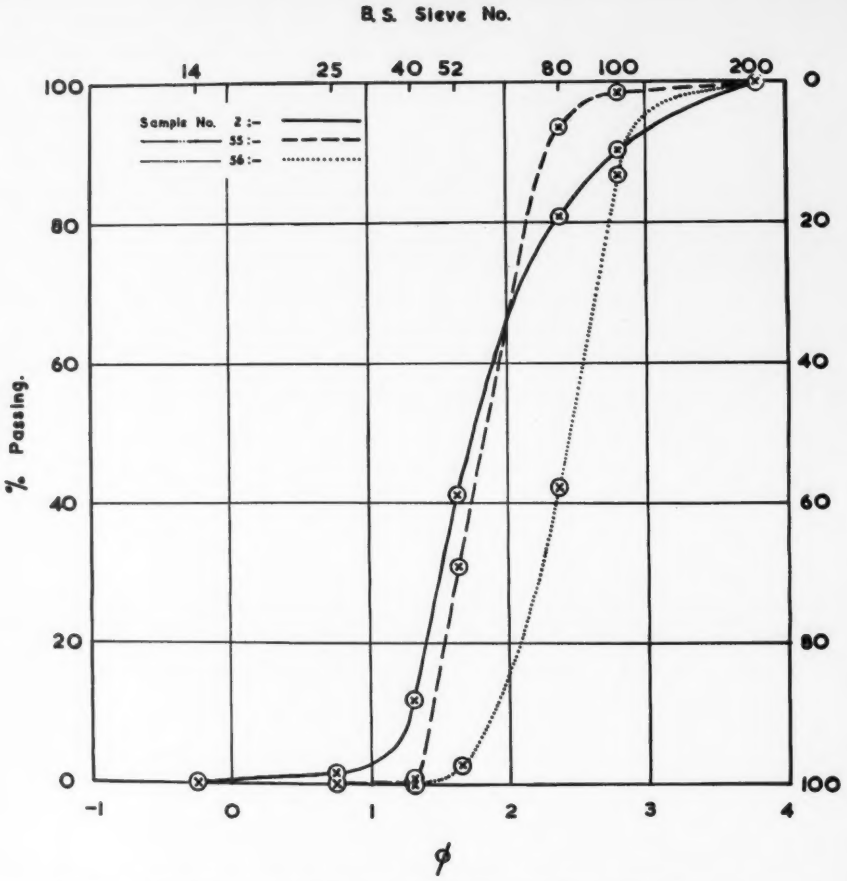

FIG. 6.—Cumulative curves for the three samples from the shore profile at Sandy Cove.

found on the shallow marine flat (which is always submerged), but the percentage is too small to affect the results or to be determined with any degree of accuracy. They are, of course, absent on the erosion surface (refer to the Bir Themada profile). All three samples have the same minimum grain size, suggesting that the differences in grain size distribution are caused by sorting, not by breaking and smashing of the parent grains.

In detail, the material actually being eroded and sorted on the beach slope differs from the parent material by:

(1) Better grading.
(2) Reduction in the quantity of material in the finer extra-modal curve.
(3) A decrease in maximum grain size.
(4) Reduction of the coarser extramodal curve to practically nothing.

The material forming the marine flat differs from that forming the erosion surface by—

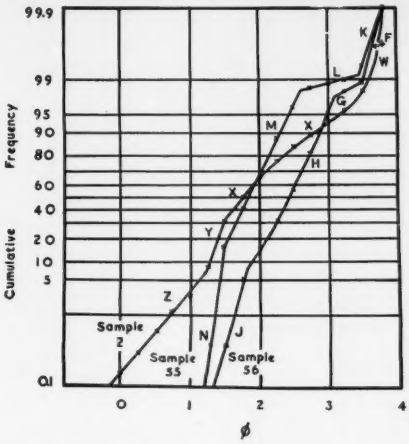

FIG. 7.—Logarithmic probability diagram for the samples from the Sandy Cove shore profile.

(1) A reduction in the maximum grain size.
(2) A lateral shift of the median.
(3) A slight increase in the coarser extramodal curve.
(4) A reduction in finer extramodal curve.
(5) Slightly poorer grading.

The best grading is therefore found on the erosion surface, with an overall reduction in grain size and sorting immediately offshore. The fluctuation in the height and wavelength of the waves probably increases the erosion occurring by providing a mechanism for the removal of the material being sorted by the waves. It is rarely that the waves break on the shore with a non-directional effect, and such a condition will only continue for as long as the wind continues to blow from that one particular direction. In any other case, longshore drift will occur as a consequence of the oblique direction of movement of the waves.

The results of plotting samples 2, 55, and 56 on logarithmic probability paper will be found in figure 7 and table 3. All follow a logarithmic probability function, although sample 2 shows modifying effects on the finer grades.

In sample 2, the third line shows very good grading but low quantity, while a small amount of coarser material is present due to the action of surface creep. The bulk of the deposit is made up of the interstitial material which includes exceptionally coarse grades. This is probably due to deposition in water.

Sample 55 includes a very high percentage within the third line, but it is less well graded. However, the limits of grain size of each line have been drastically altered, only the minimum grain size remaining constant, indicating little or no abrasion. In this case, the degree of sorting of the material moved by surface creep is much higher than for the Bir Themada samples, while the phi limits of each line are markedly different. The quantity and degree of sorting of the interstitial material is very low (again, refer to Bir Themada).

Sample 56 is similar, but there is once again a decrease in grain size in all except the first (finest) line, relative to sample 55. An increase also occurs in the quantity occurring in the third line at the expense of its neighbours, while the corresponding value of ρ also increases slightly (c.f. the overall grading). The effects of surface creep in zone DB can be seen by the presence of a moderate fourth line, which agrees with the field observations of the disturbance of the shallow surface of the mud flat by the waves.

Hence the orbital motion of the waves which operates on the beach face and on the surface of the sea bottom in the zone immediately prior to the position of commencement of the breaking of waves would appear to be responsible for the occurrence of the fourth line (representing material moved by surface creep) in the near tidal deposits. Other tidal currents do not appear to be strong enough, at any rate in the cases studied above, to cause surface creep. It may well be that the necessary turbulence is lacking.

The results of graphical comparison of median and log. E.G. for the samples are shown in figure 8. The sorting on the beach face produces a marked coarsening of median yet only a small coarsening of log. E.G., bringing the material to a position almost resting on the line corresponding to median = log. E.G. This reflects the increase in perfection of grading and the slight decrease in grain size of the modal curve.

Upon the commencement of travel from the erosion surface the grain size rapidly decreases, with the composition of the material slowly crossing the median = log. E.G. line to come to lie on the median side of it.

The result of plotting graphically the ranges of grades coarser and finer than the log. E.G. is even more revealing. The effect of the sorting on the erosion surface is seen to manifest itself in the reduction in the range of coarser grades with a negligible effect on the finer grades. However, the change taking place during transport is exactly the opposite, only the finer grades decreasing. Thus the slight increase in the coarser extramodal curve is balanced by the poorer grading.

This beach profile demonstrates the type found where the tidal influences and offshore currents are non-existent, that is where only wave action is involved. It also agrees with and explains the observations of Carroll (1939) who claimed that the wave action and not the actual tide was responsible for the sorting carried out on the marine beach face.

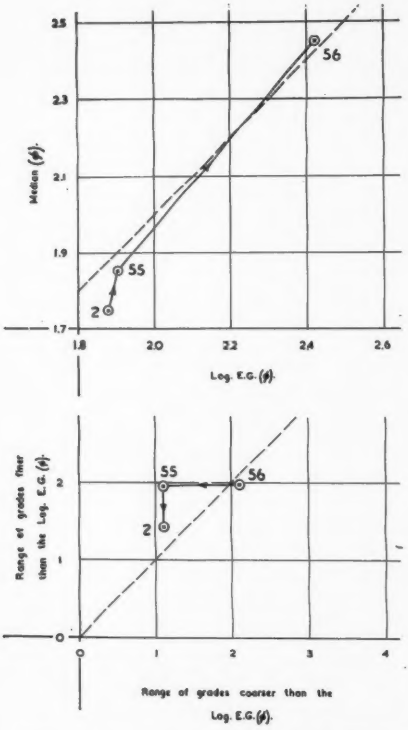

FIG. 8.—Graphs showing the relationship between median and logarithmic equivalent grade, and the grain-size range coarser and finer than the logarithmic equivalent grade for the samples from the Sandy Cove profile, Suez Canal Zone.

MECHANICAL COMPOSITION OF BEACH SANDS AND GRAVELS

Various other studies of beach profiles were carried out by the writer in the west of England (Harris, 1955). All gave similar results to the Bir Themada profile, although slight anomalies were encountered in cases of shallow shelving beaches, *e.g.* at Croyde Bay near Barnstaple, North Devon. Changes in height of the successive tides caused zone C to occur in different places, giving a succession of belts with coarser textured deposits.

The nature of the parent material must, however, be considered when studying the processes occurring on a beach. This is very well shown by the samples 58 (beach sand) and 57 (parent wadi gravel) from the shores of the Dead Sea at Kallia in Palestine (table 2). Shape of the cumulative curve is certainly not constant for beach sands derived from different parent materials. However, the same increase in degree of sorting and removal of the finest material occurs during wave action regardless of parent material. Studies of the component curves are of little diagnostic value apart from ensuring that the samples were adjusted to their depositional environment. This was shown by Emery (1955, p. 48). Beach sands may have a far greater range of grain size than dune sands, so that our attention must therefore be concentrated on the deposits having a similar range of grades.

Graphs of median against logarithmic equivalent grade are far more successful. figure 9 shows the field of median/log.E.G. for all the samples from zone DB studied by the writer (table 4). Thirteen samples of beach sand were chosen which lie within the same range as aeolian sands of the full possible range of grades. They lie within a narrow

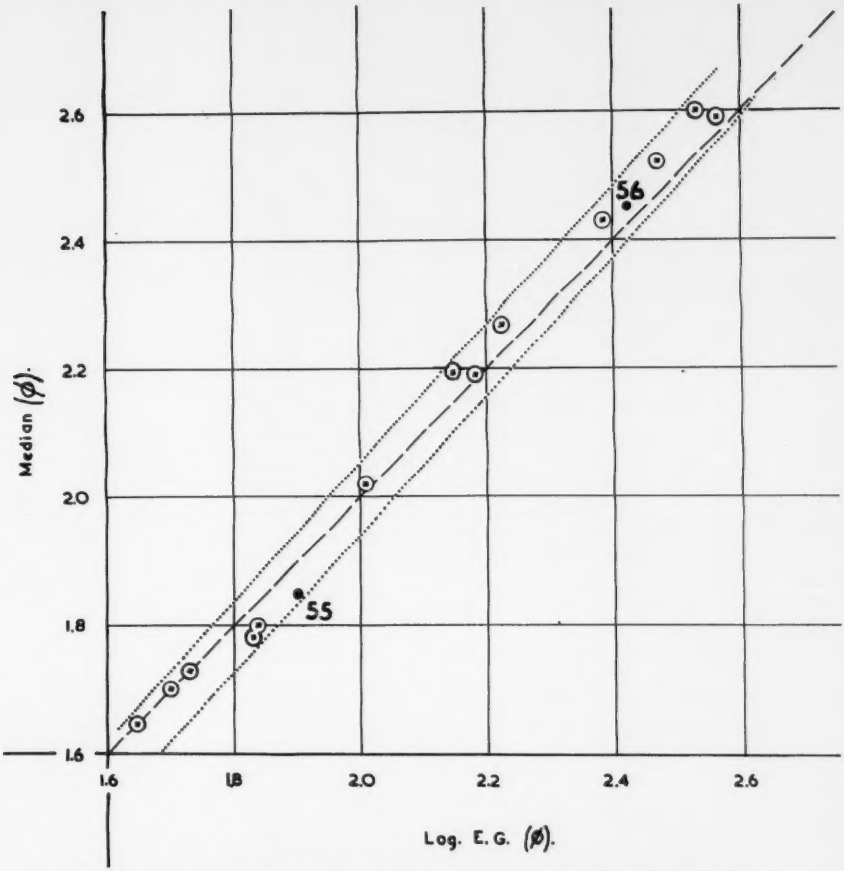

FIG. 9.—Graph showing the relationship between median and logarithmic equivalent grade for all the samples from zone DB, studied in Harris, 1955.

belt which is not quite parallel to the line corresponding to log. E.G. = median. The samples showing the effect of wave motion with and without transport (55 and 56 from Sandy Cove, Suez Canal Zone) also fall in this field. In the grades finer than about 2.00/2.00 ϕ, the median usually exceeds the log. E.G., whereas on the coarser side, the reverse occurs. This might be due to chance sampling, but it should be remembered that samples from both Egypt and England are involved, and that those from the latter come from a wide range of environments, including different source materials scattered over a distance of 50 miles of coastline. Thus it might be expected that these samples would be fairly typical.

A possible reason for this is that the finer grades tend to be removed, but the coarser grades are changed by wave action to a lesser extent in the range examined here. Undoubtedly a wider degree of scatter is to be expected as wider ranges of coarser grades (and hence a coarser median) are

TABLE 4.—*Some examples of mechanical composition of beach sands from zone DB from various localities in south-west England*

		Sample nos.										
		111	112	113	114	115	118	119	120	123	134	135
Percent passing B.S. sieve no.	$\frac{3}{8}$″	—	—	—	100	—	—	100	—	—	—	—
	$\frac{3}{16}$″	—	—	—	99.9	—	—	99.95	—	—	—	—
	7	—	100	—	99.9	—	—	99.95	—	—	—	—
	14	—	99.95	100	99.7	—	—	99.9	—	—	—	—
	18	100	99.8	99.95	99.65	—	100	99.8	—	—	—	—
	25	99.9	99.6	99.9	99.6	—	99.9	99.7	—	—	100	—
	36	99.7	99.2	99.85	99.55	100	99.8	99.4	100	100	99.9	100
	52	95.9	93.9	98.6	99.1	99.2	94.5	81.6	99.5	78.2	98.5	99.9
	60	75.3	71.3	88.3	90.4	91.7	15.0	17.5	50.6	11.4	63.7	88.3
	85	26.1	31.4	40.8	51.4	61.7	1.0	1.4	1.1	0.2	27.4	55.2
	100	11.1	17.7	21.6	33.4	43.0	0.4	0.7	0.4	0.1	17.1	46.2
	150	0.1	0.0	0.0	0.3	0.3	0.0	0.0	0.0	0.0	0.2	0.7
	200	0.0	—	—	0.05	0.0	—	—	—	—	0.1	0.1
	300	—	—	—	0.0	—	—	—	—	—	0.0	0.0
Grain size (ϕ)	Median	2.27	2.27	2.43	2.52	2.60	1.73	1.70	2.02	1.645	2.19	2.59
	Log.E.G.	2.225	2.225	2.385	2.47	2.53	1.73	1.70	2.01	1.647	2.155	2.565

111–113 From Dalmer Bay, near Padstow, North Cornwall.
114–115 From Instow, near Barnstaple, North Devon.
118–120, 123 From Croyde Bay, near Barnstaple, North Devon.
134–5 From Saunton Sands, near Barnstaple, North Devon.
See also samples 53–56 from Egypt.

present in the parent material, since sorting is so slow. However, a greater range of finer material (and hence a finer median) is likely to be of limited effect, since the finest grades are so easily sorted over and removed in suspension. By this reasoning, the transition between these two regimes where sorting of both ends of the distribution are of equal importance (i.e. log E.G. = median) is at a median value of 2.0 ϕ

Ranges of grades coarser and finer than the logarithmic equivalent grade at each site tended to fall into a long narrow field such as would be obtained by including samples 51 and 52 with samples 53 and 54 of figure 4 in a single field. The only dune sand microenvironment having this shape was the one corresponding to wind erosion of parent material. Since the shape of the log. E.G./median graph is rather different over the limited range found in aeolian deposits with a full range of possible grades, it would seem likely that these two environments can be differentiated using mechanical analysis alone.

The probable sorting processes occurring in the various zones are:

Zone A—A source of parent material, obtained by the processes of wave erosion at the base of the cliff.

Zone B—The effect of the sorting action of waves during equinoxial and freak storms is demonstrated.

Zone C—Residual deposit found where the waves have been breaking for a sufficiently long period at the same level or at sea level when the tide is rising. This leaves a pile of material too coarse to be removed by the weak undertow from where the wash has thrown it. Further down the slope, the undertow includes increasing amounts of water moving faster, with consequently much more transporting power (Steers, 1953, p. 12). The finer material is small enough to be removed by the weak currents of the undertow, hence the pebble concentrate. It is commonly found marking the high tide mark, and the process of its formation is probably the same as that resulting in the shoreward movement of spits and bars.

Zone DA—This is the surface on which the waves are breaking and sorting over the material under normal conditions. Since the tide causes rise and fall of this zone of sorting, the consistency of sorting over most of its surface is achieved, while this is also the explanation of the trend of the contours of the medians, etc., parallel to the length of the beach, noted by Krumbein (1938), and Krumbein & Griffith (1938).

Zone DB—This is a zone found where accretion is occurring and the zone of sorting by wave action on the beach slope is less than the intertidal zone. This appears most commonly to occur in estuaries or inlets where the tidal range is unusually large and where marine currents are present but fail to remove the vast quantities of sediment as quickly as it is deposited by the local rivers and streams.

Zone E–H—Below the low water mark, the material removed by the processes affecting the zones considered above will tend to be deposited as far as the currents allow, hence the reappearance of the coarsest and finest grades in the wholly marine zones interpolated from McKelvey's histograms. This also explains the variation in these zones from place to place, since different currents occurring in different areas would cause the presence of different series of zones. In addition, distribution and initial grain size of the parent source material will affect the number of possible zones regardless of whether similar currents occur in the two areas or not. However, most writers agree that a general correlation of current strength with grain size and degree of sorting may be found in wholly marine deposits on continental shelves.

ACKNOWLEDGMENTS

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FIGS. 1–2, SEPTEMBER, 1959

EVIDENCE OF WEATHERING AT THE SILURIAN-DEVONIAN CONTACT IN CENTRAL OHIO[1]

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ABSTRACT

Preservation of a lens of Hillsboro-like sandstone at the Silurian-Devonian disconformable contact in Madison County, Ohio, coincides with the presence of a dolomite pebble basal conglomerate with a red clayey matrix. Analysis of the minerals in the red matrix and study of the features of the underlying dolomite suggest that a portion of a pre-Devonian weathered surface has been preserved. The presence of such a clay mineral assemblage in other similar geologic circumstances may provide a criterion for the identification of an unconformity.

The contact of Silurian rocks with Devonian rocks is exposed in several localities in western Ohio, and study of the details of the various expressions of this contact has contributed much to the understanding of the stratigraphy and geologic history of the area. A recently exposed portion of this disconformity in a quarry in Madison County in central Ohio presents some added information concerning the stratigraphic relationship, the distribution of the Hillsboro-Sylvania sandstone, and the record of post-Silurian weathering as seen in a remnant of a fossil regolith.

The Madison Stone Company quarry is located on the north bank of Little Darby Creek in the middle of the east edge of Madison County, about three miles northwest of Georgesville and two miles southeast of West Jefferson (London quadrangle). The quarry is in a cut-rock terrace of dolomite and limestone covered with two to four feet of drift. A red residual soil which lies between the drift and the underlying limestone fills solution cavities and widened fractures (Summerson, 1959, p. 430). About midway in the twenty feet of bedrock exposed in the quarry is the Silurian-Devonian contact.

From the base upward, the section includes the following units:

1) Bass Islands Dolomite—ten to twelve feet of argillaceous, buff, fine-grained dolomite in beds that vary from thinly laminated units of two to four inches in thickness to massive beds up to three feet thick. Black and white banded chert nodules are present in regular layers one to two inches thick. In places, portions of the thinly laminated beds show a brecciated structure. Fossils, though scarce, are indicative of Late Silurian age but are not characteristic of any particular formation of the Bass Islands group. Lithologically the dolomite is distinguishable from the Tymochtee Formation but is similar to both the Raisin River Formation and the Put-in-Bay Formation. Carman (1927) has stated that the Silurian rocks in this area are in the Bass Islands group above the Tymochtee and, because of the absence of the strong brecciation characteristic of the Put-in-Bay Formation, are probably part of the Raisin River Formation.

2) Disconformity—the contact between Silurian and Devonian. In various parts of Ohio, the stratigraphic units differ on both sides of the unconformity. The underlying unit may be as old as Niagaran or as young as the uppermost Bass Islands group, of Cayugan age. The overlying unit may range stratigraphically from the Sylvania Sandstone to the Ohio Shale and lithologically from limestone or dolomite to sandy limestone or sandy dolomite, to orthoquartzite, or, in southern Ohio, to black shale. Sporadically the basal part may be represented by a conglomeratic layer containing pebbles of the underlying Silurian rocks.

In the Madison quarry exposure, the contact varies. In the western part it is a smooth, sharp, somewhat wavy line between slightly sandy dolomite of the Columbus Limestone above and buff argillaceous dolomite of the Bass Islands group below. In the

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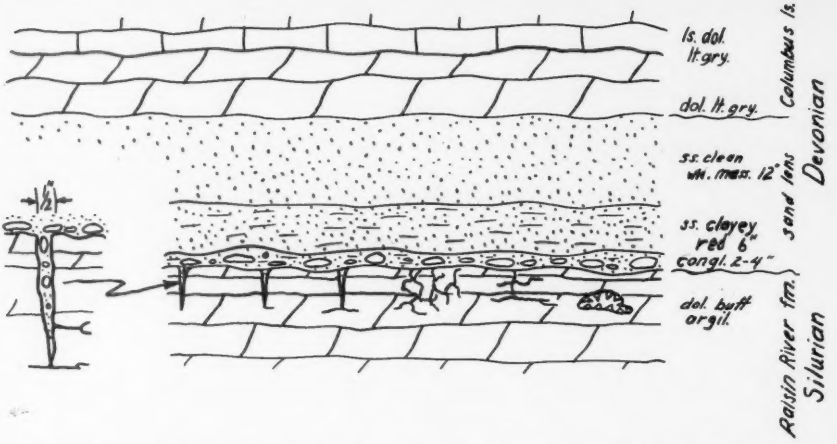

FIG. 1.—Generalized section across the Silurian-Devonian disconformity at thickest development of sand lens in Madison quarry. Conglomerate-filled fractures, clay stringers, and vugs shown in position, but with scale exaggerated.

northeastern part of the quarry a lens of sandstone and conglomerate as much as two feet thick separates the Columbus and Bass Islands dolomites.

3) Columbus Limestone—seven to ten feet of light gray to light buff, medium-grained calcareous dolomite in beds from four to ten inches thick, very fossiliferous and with some scattered white chert nodules. Upward, the formation becomes more calcareous and fossiliferous. The base is the sharp wavy line of the disconformity, and the top is the glacially eroded surface overlain by drift. The basal foot of the Columbus Limestone is somewhat brecciated, with angular and rounded fragments of the same general lithology.

The Columbus Limestone of the Madison quarry lacks the brown, massive, saccharoidal lithology typical of the lower Columbus, Bellepoint Member, to the north and east. Instead, the lithology and fauna more closely resemble the upper Columbus, or Delhi Member. If this correlation is correct, the upper Columbus overlaps the lower to the south and west of the type area.

4) Sandstone lens—Of the various features of the disconformable contact in the Madison quarry, the sandstone lens (fig. 1) is of particular interest. Occupying the position of the Hillsboro-Sylvania Sandstones, it includes two to four inches of conglomerate, four to six inches of red clay and iron-cemented sandstone, and fifteen to eighteen inches of clean, white, well-sorted quartz sandstone. The whole unit pinches out within a distance of one hundred feet. The conglomerate contains half-inch to four-inch well-rounded pebbles of dolomite with a few pebbles of chert, all of lithologies which occur in the underlying Silurian rocks. The matrix of the conglomerate is brownish-red and consists of a mixture of clay, sand, some dolomite granules, and strong ferruginous cement. The pebbles of the conglomerate are quite similar to those of the basal conglomerate of the Columbus Limestone elsewhere as described by Stauffer (1909) and of the Sylvania Sandstone as described by Carman (1936). The conglomeratic layer here differs in having a large quantity of clay and iron in the matrix. The overlying red sandstone consists of well-rounded, well-sorted quartz grains and is cemented with clay and iron oxide, which tend to decrease upward. The uppermost unit is a clean, white, friable, well-sorted, massive orthoquartzite. The quartz grains are a mixture of subangular and recrystallized grains, the latter showing prominent secondary crystal

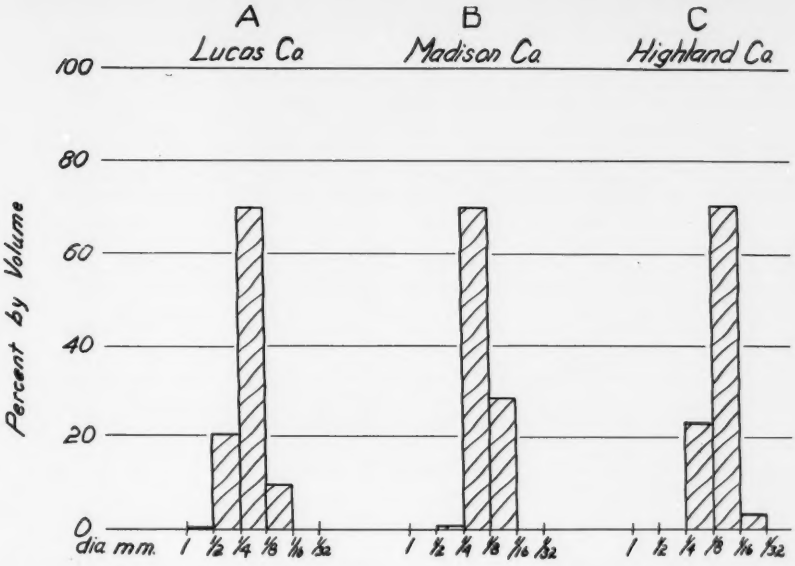

FIG. 2.—Histograms comparing size distribution of A, Sylvania sandstone, Lucas County, Ohio; B, Madison quarry lens, Madison County, Ohio; C, Hillsboro sandstone, Highland County, Ohio. Histograms A and C taken from Carman (1936).

faces on older sub-rounded and somewhat frosted grains. Mechanical analysis of the sand is shown by the histogram in figure 2B, and the size distribution is compared with that of the Hillsboro and Sylvania, as shown by histograms (fig. 2 A and C) taken from Carman (1936). When adjustments of grain size are made to compensate for the secondary enlargements, the size distribution of the sand lens is intermediate between that of the Sylvania and the Hillsboro.

Contrary to Ulrich's (1911) observations of an absence of sand at the systemic boundary in the Cincinnati Dome area, one of the most persistent markers of the Silurian-Devonian contact in Ohio is the presence, in varying amounts, of round, frosted, quartz sand grains. Locally the sand is thick enough to be identified as a formation (e.g., the Hillsboro Sandstone in southern Ohio and the Sylvania Sandstone in northwestern Ohio); elsewhere, it is a sandy zone at the base of the Devonian. Detailed description of this sandy zone has been given by Summerson and others (1957). In his paper on the Sylvania Sandstone, Carman (1936) has discussed the possible eolian origin and stratigraphic position of this sand. Except for the strong recrystallization of the sand grains, the lens at Madison quarry is identical to the Hillsboro. It is interpreted to be a local concentration of the sand sheet common to the Silurian surface, preserved in a depression during the advance of the Columbus sea.

In the area of the sand lens, the Silurian rocks show several structures that suggest the weathering conditions of pre-Devonian time. In places the Silurian rocks have fractures one-half inch wide and from four to six inches deep in irregular polygonal patterns, filled with the red conglomerate. In other places the top part of the Silurian dolomite consists of irregular angular fragments with thin, vein-like stringers of red clayey cement between them. Other features along the surface include vugs lined with calcite crystals and coarsely recrystallized

areas of calcite and dolomite, with some of the calcite of a dark brown variety. All of these features, unique to the area of the sandstone lens, are suggestive of ground water action, and contrast with the usual smoothly planed surface of the Silurian elsewhere.

Because the red, clayey matrix of the basal conglomerate is unusual, analyses were made of it and the underlying dolomite. X-ray diffraction studies were made of the finer portion of the insoluble residues of the red matrix, and the following minerals were identified: kaolinite, illite, mixed layered material, goethite, and quartz. Analysis of the rock below the disconformity showed it it to be dominantly dolomite, with minor amounts of calcite and quartz. Analysis of the insoluble residues of the dolomite showed the presence of illite, quartz, and possibly some kaolinite.

In his discussion of criteria for the recognition of unconformities, Krumbein (1942) has suggested the study of chemical and mineral variations in the vicinity of the contact as a means of identifying a buried soil profile, e.g., a concentration of iron or a particular clay complex. In a study of weathering of clay-size particles, Jackson and others (1948) determined a sequence of minerals formed as weathering progresses and the association of minerals that would be present at each stage. Jackson's findings may be applied to the study of the unconformity in the Madison quarry to interpret the origin of the red matrix as being from the underlying bedrock. Because the Silurian dolomite underlying the unconformity contains illite, soil formed from this dolomite would contain clay minerals that could have been developed by weathering processes. The matrix of the conglomerate is composed of clay minerals, dominantly kaolinite, and goethite. Together, the kaolinite and goethite represent stable end products of weathering alteration, as pointed out by Frederickson (1952). According to Jackson's table of weathering sequence for clay-size minerals, the development of a kaolinite stage would follow that of an illite stage. Likewise, the association of minerals present in the matrix is similar to that postulated by Jackson to be present in the kaolinite stage. Thus the sequence of clay alteration and the association of clay minerals suggest that the red matrix is a remnant of a mature residual soil formed by weathering of the Silurian dolomites.

Supporting the identification of the matrix of the sand lens as a remnant of a pre-Devonian fossil soil is the association of such features as the recrystallized calcite, the vugs, and the red clay-filled fractures. All of these features are limited to the area of the Silurian surface covered by the sand lens. It seems probable that this portion of the surface was a depression where the weathered rock and soil were covered by the sand and protected from the reworking, washing, and erosion by the advancing seas that occurred on nearby areas.

In summary, the exposure of the Silurian-Devonian contact in the Madison Stone Company quarry offers further information on the distribution of the sand at the contact, a record of a pre-Devonian soil, and, of particular interest, an example of the possible use of clay-size mineral associations as a criterion for the recognition of weathering and its stages along an unconformity.

ACKNOWLEDGMENTS

The following persons generously helped in this study: Mr. M. A. Pfeifer, owner of the Madison Stone Company quarry, who allowed frequent access to the quarry and the collection of samples; Dr. H. D. Glass, Illinois State Geological Survey, who made X-ray identification of the minerals and offered helpful discussion of results; and Dr. H. H. Murray, Georgia Kaolin Company, Elizabeth, New Jersey, who read the manuscript critically.

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FIG. 1, SEPTEMBER, 1959

PRE-GLACIAL RESIDUAL SOIL IN CENTRAL OHIO[1]

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ABSTRACT

Study of a cherty, red, residual soil found in Madison County, Ohio, has resulted in the reconstruction of the chemical and environmental conditions necessary for its formation and the postulation of its probable age. The soil exists in solution pits and fractures in the Columbus Limestone, overlain by Wisconsin drift. Studies of field relations and insoluble residues show the limestones to be the source rock. Soil and mineral analyses, considered on the basis of weathering necessary to alter the source minerals to minerals in the soil, indicate that the soil must have formed in a warm climate with heavy seasonal rainfall in an area of moderate relief. The relation of these conditions to the geologic history of the area suggests that the soil formation is of pre-Pleistocene age.

PRE-GLACIAL RESIDUAL SOIL IN CENTRAL OHIO

In most of the glaciated parts of Ohio the drift lies directly on slightly weathered to fresh bedrock. Efficient erosion by the glaciers has largely destroyed the record between. An exception to this condition, a red residual soil preserved beneath the Wisconsin drift, was recently found in Madison County. The following is a description of the residual soil and an interpretation of the conditions it records. The study is offered as an example of a contribution to historical geology from clay mineral and petrographic analyses.

The residual soil is exposed in the Madison Stone Company quarry near the middle of the east side of Madison County, Ohio (London quadrangle), about three miles northwest of Georgesville and two miles southeast of West Jefferson. The quarry is located in a rock terrace approximately forty feet above the north bank of Little Darby Creek. The terrace was shaped by glacial erosion which removed most of the soil and planed the bedrock to a generally level surface. The bedrock section includes about eleven feet of buff, fine-grained, argillaceous dolomite of the Bass Islands group (Silurian) disconformably overlain by seven to ten feet of light gray to buff, medium-grained, slightly cherty dolomitic limestone of the Columbus Formation (Devonian). The generally level but slightly undulating surface of the limestone is overlain by as much as two feet of sand and gravel which in turn is overlain by a gray-brown till as much as two feet thick in the area of the terrace and thickening toward the valley wall to the east (fig. 1). The till is a part of the Scioto lobe (Late Wisconsin—the nearest radio carbon date of ice advance is 23,000 years) deposits that cover much of Franklin County, as mapped by Goldthwait[2] (personal communication). The bedrock, stripped in preparation for quarrying, is interrupted by numerous solution-widened fractures and occasional solution pits filled with the red residual soil. Together, the bedrock and residual soil form a smooth surface; the contact with the overlying drift is continuous and sharp.

Some of the residual, soil-filled depressions and cavities are funnel shaped, and are two to five feet deep and two to three feet wide. Others are elongate fractures, three to ten inches wide, which extend downward from ten to twelve feet into the bedrock, forming irregular fissures, branching and enlarging into cavities of various shapes; these are up to three feet or more in diameter; all are filled with the cherty, red, residual soil. Quarrying operations, at different times, have exposed a residual soil-filled channel eight to ten feet wide and four to five feet deep, and a fossil sink about twenty feet wide and fifteen feet deep. The sink contained, in addition to a fill of red residual soil, some blocks of limestone in jumbled positions, with some Columbus Limestone blocks below the Bass Islands contact; the contact with the overlying drift showed a slight depression but no recognizable mixing.

[1] Manuscript received January 19, 1959.

[2] Prof. R. P. Goldthwait, Department of Geology, Ohio State University.

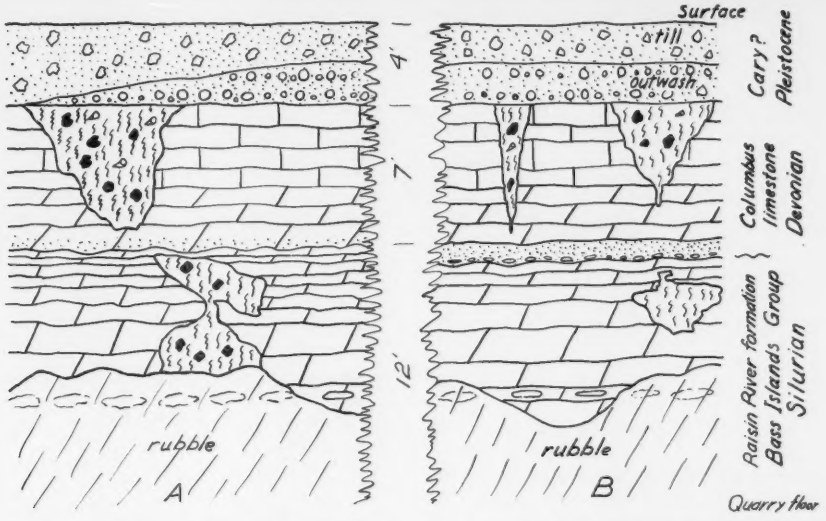

FIG. 1.—Generalized sketches of quarry wall showing positions of residual soil-filled cavities and fractures. A, section in west wall; B, section in east wall. Samples 1 and 2 were taken from near center of a depression filling and a fissure filling, respectively.

A red residual soil similar to that in the Madison quarry was recorded from the Marble Cliff quarries near Columbus, Ohio, by Hubbard in 1911 (p. 62–63). Other exposures of similar red residual soils have been noted recently by Holowaychuk[3] and Forsyth[4] (personal communication) in the vicinity of Springfield, Ohio, and in southern Fayette County. Caves in the region—e.g., Olentangy Caverns northwest of Columbus—characteristically have portions filled with red mud, and the walls of the open passages are heavily coated with similar mud.

DESCRIPTION OF THE RESIDUAL SOIL

The soil is deep red (10 R 4/6 on the Rock Color Chart), predominantly clay, with some silt and sand. Results of mechanical analyses are shown in table 1.

Pebble-size fragments include chert nodules and nodule fragments and the larger silica-replaced fossils, chiefly corals. Smaller fragments, granule to small pebble size, consist of silica-replaced brachiopods and crinoid and blastoid plates, masses of drusy quartz, and chalcedonic spherulites and oölites. Many of the fossils show a beekite structure common in chalcedonic fossil replacement. Material of coarse sand size includes oölites of chalcedony and well-rounded frosted quartz grains. Fine sand and silt sizes include, in the coarser part, angular quartz grains and, in the finer, both quartz and chalcedonic fragments. The sand and silt sizes are characterized by bimodal distribution, the coarser being 1/4–1/32 mm and the finer 1/64–1/256 mm. The clay size fraction was studied by X-ray diffraction.

The soil has a pH of 7.1, a base saturation of 89 percent, and a base exchange of 36 me/100 g. Its red color and high clay content suggest that this soil in its original

TABLE 1.—*Mechanical composition of red soil*

	Size range	Sample 1	Sample 2
Sand	.05 mm–2.0 mm	14.3%	11.2%
Silt	.002–.15 mm	20.4%	20.9%
Clay	smaller than .002 mm	65.3%	68.0%

[3] Holowaychuk, Nicholas, Prof., Department of Agronomy, Ohio State University.

[4] Forsyth, Jane L., Geologist, Ohio Geological Survey.

TABLE 2.—*Clay mineral composition:* percentages calculated on the basis of the total original sample[1]

Sample	Quartz	Kaolinite	Illite	Montmorillonite
1	20	11	9	11
2	45	5.5	6.5	4
3	22	7	12	13

[1] Analyses were made at the Jersey Production Research Company through the courtesy of Dr. A. L. Kidwell.

state was similar to the Reddish-Brown Latosolic or the Red-Yellow Podsolic soils developed on limestones in south central United States and to the "terra rossa" of the Mediterranean area.

X-ray diffraction studies were made of the minus 2 micron fraction. Samples were prepared by disaggregation in a Waring blendor and the 2-micron fraction was separated by centrifugation. In order to make more precise identification of the clay minerals, oriented aggregates were prepared by use of a porous plate technique. The samples were glycolated to aid in identifying montmorillonite and heated to distinguish kaolinite from chlorite. Results are shown in table 2. Iron was identified by polarograph. The quantities from three samples are shown in table 3. The presence of large percentages of iron gave high background and caused difficulty in identification of the randomly oriented samples. Treatment of some of the samples with oxalic acid caused the removal of as much as 50 percent of the sample. However, not all of the dissolved material was iron; some was probably amorphous material and some probably montmorillonite. Whereas both goethite and hematite are indicated on some of the diffraction curves, the distinctly red color of the soil suggests that most of the iron is in the form of hematite.

TABLE 3.—*Iron content*[1]

Sample	Percentage of Fe
1	10.06
2	3.87
3	4.00

[1] Analyses were made at the Jersey Production Research Company through the courtesy of Dr. A. L. Kidwell.

RELATION OF SOIL TO BEDROCK

From the field relations, the soil is judged to be derived directly from the weathering products of the surface bedrock, the Columbus Limestone. All of the coarser constituents of the soil—chert nodules, silicified fossils, chalcedonic oölites and spherulites, and detrital quartz grains—are identical with the material in the insoluble residues of the Columbus Limestone as described recently by Summerson and others (1957).

With only the lower eight to ten feet of the Columbus Limestone present (the original thickness was sixty feet or more), it is unlikely that the soil in the fracture filling, representing the lowest part of the original residual soil profile, would contain any material other than that derived from the Columbus Limestone. Nor is there any evidence of contamination from mixture with the overlying drift; the contact is sharp and clean.

Analyses of the Columbus Limestone show variation in composition from dolomitic limestone to calcareous dolomite, with the dolomite more dominant toward the base. X-ray diffraction curves show, in addition to the carbonates, a recognizable peak for quartz. Diffraction curves of the insoluble residues show quartz and illite. The amount of illite is approximately one percent of the original sample. Insoluble residues of the Columbus Limestone typically contain from 1 to 3 percent of clay and silt-size particles and ½ of one percent to 2 percent of sand-size particles.

Evidence from both the field relations and the comparison of the insoluble residues of the limestones with the coarser material of the soil indicates that the limestone is the source or parent rock of the soil. The most striking difference is in the clay minerals: the clay of the limestone is illite, whereas that of the soil is a combination of illite, montmorillonite, and kaolinite.

INTERPRETATION OF THE SOIL RECORD

Occurrence and distribution of clay minerals in sedimentary rocks and soils have been described by a number of authors. Van Houten (1953) has discussed the influence of source material on the distribution of clay minerals in soils and has pointed out that the source rock controls the distribution of

the hydrous micas characterizing many soils. Ross and Hendricks (1945), Grim (1953), Keller (1956), and others have described many examples of clay mineral associations in soils, emphasizing alteration by weathering to meet new physical and chemical conditions of the environment. The soil described here illustrates the importance of both controls and demonstrates the influence of the sedimentary rock source of the clay minerals and the development of new minerals in response to a change in environment.

If the evidence for the Columbus Limestone as the source of the residual soil found exposed in the Madison quarry is accepted, it follows that the source of the clay minerals is probably the illite present in the limestone, which is freed to the soil upon solution of the parent rock. Thus montmorillonite and kaolinite in the soil may be alteration products of the weathering actions on the illite. Jackson and others (1948 and 1953), in papers on the weathering sequence of clay-size materials, have shown the minerals and the associations of minerals that may be expected to be present at any one stage in the weathering development. In the 1948 paper Jackson suggests that such an association occurs in the quantities of a normal distribution curve, with one or two of the minerals dominant (40–60 percent). Whereas the proportions of the minerals in the Madison quarry residual soil are not those postulated by Jackson in his distribution curve, the same association of minerals is present for the corresponding weathering stage. The variations in the proportions from the suggested curve may be explained by two conditions of the residual soil: first, since the major part of the soil was removed by the glaciers, the remnant of residual soil would represent the lowest part of the soil profile, the portion of least alteration; second, the lower portion of the soil is also in a position to receive the steady influx of a new supply of illite from the continuing solution of the source limestone.

For comparison of the clay mineral development in similar limestone soils, the findings of Murray and Leininger (1956) and Carroll and Hathaway (1953) are of particular interest. Murray and Leininger made a study in Indiana of the soils on Wisconsin and Illinoian tills and of the residual soil on the limestones south of the glacial border. The latter, the limestone soil, is described as the mature stage or oldest in the sequence of clay mineral alteration shown by the three soils. The Indiana limestone soil is a red-brown podsol characterized by a dominance of montmorillonite, with minor amounts of chlorite, kaolinite, illite, and mixed layer materials. The montmorillonite stage of the Indiana soil is either younger than the Madison quarry residual soil or represents a condition of less intensity of weathering and/or a more alkaline environment. The analysis of the Indiana soil does not include a description of the iron content, but its strong red color indicates considerable oxidation. Compared with this, the Ohio soil, having a greater amount of kaolinite, would seem to represent a more advanced stage of maturity. Because the sample is from the lowest part of the profile, soil from the upper portion would have included still greater quantities of kaolinite, i.e., have been even more mature.

In their study of the Lenoir Limestone soil of Viriginia, Carroll and Hathaway (1953) discussed in some detail the development of a soil similar to the Ohio residual soil. The Lenoir soil is a gray-brown podsol, becoming yellow-brown in the lower part of the profile, and containing some sand and chert fragments. Hydrous mica (illite) and montmorillonite were identified in the source rock; kaolinite and minor amounts of chlorite were the weathered products. Chemical and mineral changes involved in the weathering processes were discussed in the paper, and some estimates were made of the length of time required for the formation of the soil. Frederickson (1952) has proposed that kaolinite represents a rather stable end product in the alteration of clay minerals by weathering. In the Lenoir study the kaolinite stage is identified as the end product in the mature development of the soil. The upper part of a reconstructed profile of the Ohio Madison quarry soil would show a similar dominance of kaolinite. Thus the Ohio soil is in a weathering stage comparable to that of the Lenoir soil, but the yellow color and the clay minerals of the Lenoir suggest a stronger acidic and weaker oxidation environment.

The strong red color of the Ohio soil and results of the polarograph analyses suggest that iron in the form of hematite is an important part of the soil. The source is probably a combination of dispersed iron from the carbonate and from the illitic clay. The hematite form of iron gives further evidence as to the nature of the environment in which the soil was formed. Although the explanation of strong red color in sediments is still much debated, hematite is a rather stable end product in the weathering sequence; and most soils students tend to emphasize such conditions as strong oxidation and warm temperature for its formation. As the iron is freed from the limestone, conditions must be such that it is rapidly oxidized before it can be removed in solution. Once in the form of hematite it would tend to coat particles of quartz and clays and remain relatively stable.

From the description of the Ohio soil and the comparison with similar limestone soils it is possible to reconstruct briefly the conditions responsible for its development. The Ohio soil represents conditions of soil formation that mark a transition to the conditions of latosolization, a term coined by Jackson and others (1953) to avoid the implication of extremes in the term laterization. The formation of the Ohio soil required chemical conditions of strong leaching and rapid oxidation. The climate, therefore, must have been warm and moist. The strong leaching would result from heavy rainfall on moderate slopes, affording good drainage and allowing little accumulation of organic material. The rainfall might have had a strong seasonal distribution, with intervening dry periods tending to limit the vegetative cover and to increase the oxidation of both the organic material and the iron. In comparison with the Indiana limestone soils, the weathering of the Ohio soil seems to have been of greater intensity and of longer duration. As compared with the Virginia soils, the Ohio soil seems to represent conditions primarily of greater oxidation and also, possibly, of greater rainfall and/or greater age.

With the postulated conditions for the formation of the Ohio residual soil in mind, one can examine the geologic situation and attempt to determine the place of these conditions in the geologic history of the area. That the soil is pre-Scioto (mid-Wisconsin) is evident from its position. Its formation during the Sangamon (Illinoian-Wisconsin) interglacial interval is unlikely: the required conditions would have included a large area free of drift within the Illinoian glacial border, as well as a warm, alternating wet and dry seasonal climate, and good drainage, throughout a rather long period—a combination of conditions difficult to fit into this interval. The quantity of residual soil filling the sizable depressions indicates a considerable length of time for the development of the soil. Nor would the Yarmouth or Aftonian have been likely to meet these conditions, of which the most important would have been the lack of drift cover. Although the climatic conditions during Yarmouth time come nearest to those required for the formation of this soil, it is improbable that the area was free of drift, because it was located relatively close to the border of Kansan ice. Availability of the limestone surface to weathering as a local outcrop is also unlikely, for the area is not a bedrock high; the well-developed pattern of solution features over the area of the quarry is not indicative of a local development; and the rather widespread distribution of the reported similar soil remnants suggests a regional, rather than a local, surface development. It is not known, of course, whether the Nebraskan drift extended into central Ohio; if it did, it would have prevented the formation of the limestone residual soil at that time; if it did not, the ice was probably so close that climatic conditions were unfavorable for the formation of the soil preserved here.

If these postulates are correct, the residual soil was formed at least pre-Kansan and probably pre-Nebraskan, and therefore probably records pre-Pleistocene conditions. The last interpretation seems to fit best the conditions required for its formation.

In any further consideration of the origin of this soil, its distribution and its relationship to the solution features of the area should be studied. The several other findings of similar red residual soil should be compared with the samples described here, and study should be made of the red fills and wall coatings of caves in the region. The

presence of caves—in addition to the channeled surface, the sinks, and the fractures in the Madison quarry—suggests a widespread solution surface sometime in the past. A comparison of the conditions of this surface with Bretz's (1953) study of cave formation may add further physiographic details to the reconstruction of the period of the soil formation.

CONCLUSIONS

By studying this red residual soil of central Ohio, identifying its parent rock, and analyzing the clay-size minerals present, one can arrive at:

1. an interpretation of the chemical conditions that would account for the development of the clay-size minerals in the soil—i.e., quartz, illite, montmorillonite, kaolinite, and hematite—from the illite and quartz residues of the parent rock;

2. the reconstruction of the climatic conditions that would meet the chemical requirements, i.e., abundant rainfall, probably in wet and dry seasonal distribution, in a warm, mild climate; and

3. the probable pre-Pleistocene age of the soil and the possibility of reconstruction of the physiographic conditions of the time of its development.

ACKNOWLEDGMENTS

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FIGS. 1–6, SEPTEMBER, 1959

CALCAREOUS SPRING DEPOSITS, DUBOIS AREA, WYOMING[1]

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ABSTRACT

The thermal water deposits of the Dubois, Wyoming, area occur in five groups in the forms of domes, cliffs, flows, extinct geyser orifices, and natural bridges. The natural bridge group is not discussed. Two of the groups are along principal streams which flow perpendicular to outcrop strike; the remaining two groups parallel the strike of the outcrops.

The deposits, which are principally calcium carbonate, occur on strata ranging in age from Mississippian to Permian; the undoubted sources of the mineral enrichment were the underlying Paleozoic limestones. All of the thermal water deposits are underlain by a travertine cemented conglomerate which seemingly dates them as post-Pleistocene, but the correct age is dependent on the age of Warm Spring Canyon. The mineral bearing waters reached the surface by travelling along bedding planes and possibly along faults.

LOCATION

The Dubois area, Fremont County, Wyoming, is centered in the northwestern tip of the Wind River Basin, south of the Absaroka Mountain escarpment and north of the Wind River Range. Specifically the thermal water area begins two miles northwest of Dubois where Water Gulch enters Wind River (Sec. 2, T. 41, N., R. 107 W.) and extends northwestward to Sec. 23, T. 42 N., R. 108 W. (fig. 1).

PREVIOUS INVESTIGATIONS

The earliest work in the Dubois area was carried on by the expeditions of the U. S. Army in the period 1859–73. O. H. St. John included the area in his 1877–78 reconnaissance for the U. S. Geological Survey. During the period 1900–43 Blackwelder (1913), Miller (1936), E. B. and C. C. Branson (1937 and 1941), Love (1939), and Delo and Miner (1943) studied the geology of the Dubois and surrounding areas.

Most recent investigations include the U. S. Geological Survey Professional Paper 294-E, by William R. Keefer and the senior author's stratigraphic study of the Dubois area (Reeves, 1958). Though Keefer included much of the thermal water area in his study, "enough data were not obtained within the mapped area to explain adequately the origin of these deposists" (Keefer, 1957).

[1] Manuscript received March 9, 1959.

STRATIGRAPHY

Though the section from the Precambrian into the Tertiary is nearly complete, except of course for the absence of the Silurian, the thermal deposits are related only to the Mississippian, Pennsylvanian, and Permian strata. Therefore, only the Mississippian, Pennsylvanian and Permian strata are herein mentioned in detail.

Mississippian

Madison Formation.—In the Dubois area the Mississippian is represented by the Madison Limestone. The Madison is a blue to gray, massive bedded, medium textured crystalline limestone. Fossiliferous zones containing brachiopods, crinoids, pelecypods, and trilobites occur in the upper parts. (*See* Measured Section, Madison Formation, page 437.)

The thickness of the Madison is variable because of the unconformity at the top: Keefer (1957) quotes a thickness of 740 feet near the Livingstone Ranch in Sec. 31, T. 43 N., R. 106 W.

Pennsylvanian

Amsden Formation.—Pennsylvanian time is represented, in the Dubois area, by the Amsden Formation and the overlying Tensleep Sandstone Formation.

The basal part of the Amsden Formation is termed the Darwin Sandstone Member. Predominately of sandstone with interbedded, red, fossiliferous shales and limestones, the Darwin is often covered by

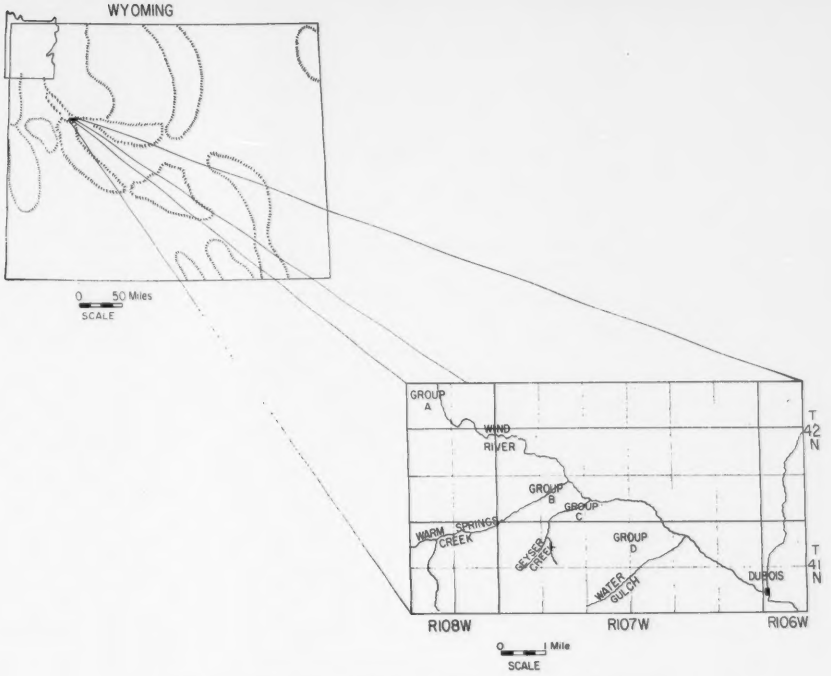

FIG. 1.—Index map showing location of thermal water area.

talus; however, along Warm Spring Canyon the Darwin tends to form a cliff.

In the southern part of the thermal water area the Darwin Member is a three-foot coarse-grained conglomerate with a matrix of coarse sand and phenocrysts of Madison limestone. This conglomerate grades upward into a light tan to brown sandstone with shale partings: total thickness of the member is 15 feet.

The immediately overlying Amsden is predominately of green and yellow arenaceous shales which rapidly grade upward into a massive-bedded tan lithographic limestone. Because the Amsden in the thermal water area is generally covered by travertine

Measured Section, Madison Formation
Sec. 1, T. 41 N., R. 108 W.

	Thickness in Feet
Amsden Formation	
Madison Formation	
3. Limestone, dark gray to black, crystalline, arenaceous lenses, fossil horizons, calcite and chert abundant	475.0
2. Limestone, reddish blue to blue, medium grained, sandy, massive, well cemented, abundant calcite	100.0
1. Conglomerage, red, angular, friable, matrix silty shale	0.5
Measured thickness Madison Formation	575.5
Darby Formation.	

Measured Section, Amsden Formation
Warm Spring Creek

	Thickness in Feet
Tensleep Formation	
Amsden Formation	
5. Sandstone, cream at bottom to red and white at top, weathers mottled red, calcareous, massive, hematite nodules	30.0
4. Sandstone, buff, fine-grained, calcareous, alternate hard and soft beds, cross-bedded	65.0
3. Sandstone, red, buff, white, fine-grained, well sorted, cross-bedded	40.0
2. Slump	5.0
1. Conglomerate, red, calcareous, sandy shale matrix, phenocrysts of fine-grained, gray limestone (Madison?)	5.0
Measured thickness Amsden Formation	145.0
Madison Formation.	

flows, the varicolors of the basal shales are attributed to oxidation of the geyser deposits.

In the thermal water area the Amsden averages 115 feet in thickness, the Darwin Member 15 feet. Keefer (1957) measured the Amsden along Little Warm Spring Creek, 3½ miles to the southeast, and recorded a thickness of 356 feet. The Darwin Member measures 29 feet east of Spring Mountain and 170 feet along Little Warm Spring Creek. These various thicknesses are no doubt the result of both the unconformity at the base of the Darwin Member and the unconformity at the top of the Amsden.

Up Warm Spring Creek and out of the immediate thermal water area the Amsden measures 145 feet thick. (*See* Measured Section, Amsden Formation.)

Tensleep Formation.—Lying unconformably on the Amsden Formation, the basal Tensleep often includes chunks of Amsden In the thermal water area the Tensleep is a 272–foot bed of sugary-textured, fine-grained, cross-bedded sand which is usually covered by travertine flows.

A partial section of the upper Tensleep is exposed in a road cut south of U. S. Highway 287, a short distance southeast of the DuNoir turnoff. The overlying Phosphoria is separted from the Tensleep by an intergradational contact between the lower sandstones and the upper limestone. (*See* Measured Section, Tensleep Formation.)

In the Dubois area the boundary between the Tensleep and the overlying Phosphoria Formation is gradational and, though rep-

Measured Section, Tensleep Formation
Dubois Area

	Thickness in Feet
Phosphoria Formation	
Tensleep Formation	
9. Limestones, brown, sandy, alternating with green shales, abundant calcite, poorly consolidated	2.0
8. Limestone and shale, alternating beds, arenaceous, buff, calcite abundant	3.5
7. Sandstone, shaly, chert nodules	6.0
6. Sandstone, green to buff, shaly, thin-bedded, chert nodules	3.5
5. Sandstone, thin beds, sandy, cherty	6.0
4. Sandstone, green to buff	7.0
3. Sandstone, buff, limonite along bedding planes, calcite geodes	7.0
2. Sandstone, white to buff, lenses of chert nodules, limonite abundant	30.0
1. Sandstone, white to buff, cross-bedded, resistant, fine to medium grained, fossiliferous near bottom: brachiopods	85.0
Measured thickness Tensleep Formation	150.0
Covered Interval	
Amsden Formation.	

resenting an unconformity of regional extent, is considered locally to be nothing more than a slight erosional surface.

Permian

Phosphoria Formation.—The Phosphoria Fomation is widespread throughout the Rocky Mountain system and lithologically is as well known, except along the southeastern flank of the Bighorn Mountains, as any stratigraphic unit.

Because the Phosphoria exhibits rapid facies changes over short distances it is usually divided lithologically into the lower, middle and upper members.

The lower Phosphiria consists of chert, sandstones, and rock phosphate which occur interbedded with shales and limestones. The middle Phosphoria is chiefly fossiliferous limestone, often cherty. The top of the middle unit is recognized by a glauconitic, phosphatic shale. The upper Phosphoria, in its lower part, is characterized by nodular, cherty limestones with thin lentils of rock phosphate; the upper part consists of a few green, glauconitic shales, glauconitic limestones and massive dolomites.

Two sections of the Phosphoria were measured in the thermal water area, one at the mouth of Warm Spring Canyon, the other at the road cut where the Tensleep section was taken. The section at the road cut is definitely lower Phosphoria, that at Warm Spring Canyon probably upper. (*See* Composite Measured Section, Phophoria Formation.)

Composite Measured Section, Phosphoria Formation
Dubois Area

	Thickness in Feet
Dinwoody Formation	
Covered	
Phosphoria Formation	
25. Limestone, light gray, massive to laminated, abundant calcite crystals, nodules of hematite	10.0
24. Phosphate, dark brown, granular	1.0
23. Limestone, black and white, mottled, chert nodules, fossiliferous: brachiopods, gastropods, bryozoans	3.0
22. Limestone, light gray, weathers light brown, massive, fossiliferous: bryozoans, brachiopods, gastropods	3.0
21. Limestone, light gray-green, weathers light brown, well jointed, chert nodules, calcite crystals	2.5
20. Shale, light gray-green, weathers dark gray, chert nodules, phosphatic, glauconitic, fossiliferous: brachiopods and gastropods	4.5
19. Sandstone, buff, massive, dense	10.0
18. Shale, blue	8.0
17. Limestone, dark gray, dolomitic, phosphatic	2.0
16. Shale, red	1.0
15. Limestone, gray-blue, dolomitic, fossiliferous: gastropods, brachiopods	15.0
14. Sandstone, white, chert nodules at bottom, chert beds at top	15.0
13. Sandstones, gray, interbedded shale with white, granular phosphate	10.0
12. Sandstone, light brown, chert, green phosphatic shales at top	10.0
11. Limestone, light brown, thin-bedded	2.0
10. Sandstone, light brown, calcareous	4.0
9. Limestone, light gray, banded cherts, limonite concertions, lenses of arenaceous limestone	11.0
8. Sandstone, light brown, calcareous, purple chert nodules	3.0
7. Limestone, gray, sandstone, brown, interbedded, chert in both	2.5
6. Sandstone, light brown to white, fine-grained	3.0
5. Sandstone, purple-white, fine-grained, calcareous, lenses of light gray sandstone	13.0
4. Shale, greenish, sandy, partly replaced by limonite, topped by light gray, calcareous sandstone	2.5
3. Sandstone, purple-brown, calcareous, calcite crystals	3.2
2. Limestone, light gray, purple chert nodules	10.0
1. Limestone, light gray to buff, laminated	12.0
Measured thickness Phosphoria Formation	161.2
Tensleep Formation.	

THERMAL DEPOSITS

A thermal water deposit results from the escape of hydrothermal waters of the Earth's crust in the form of hot springs or geysers. Resulting deposits depend on the type of rock the waters pass through before reaching the surface. If the country rock is alkaline, the deposit will be of a silicious nature such as sinter, geyserite, or siliceous tufa; however, when country rock is carbonate, resulting deposits will necessarily be of calcium carbonate. Clark (1924) has shown that when SiO_2 is the dominant compound in hot spring waters, $CaCO_3$ is never present, therefore gradational or intermittent deposits rarely occur. The thermal water deposits of the Dubois area are at least 93.6 percent calcium carbonate, with the content sometimes rising to 99.4 percent (table 1).

The classical example of thermal water deposits in Wyoming as well as the United States are those of the Yellowstone Park region. Located 60 miles northwest of the Dubois area, the thermal water deposits of Yellowstone cover an 1800 square mile area, in which Allen and Day (1935) confirmed the earlier figure of over 3600 orifices.

There are three other notable areas of thermal water deposition in Wyoming. The Stinking Water Springs located on Stinking Water River just below the canyon between Rattlesnake Ridge and Cedar Mountain; the Big Horn Hot Springs on the Bighorn River at Thermopolis; and the Fort Washakie Hot Springs situated in the valley of Little Wind River 3 miles below Fort Washakie.

The rate of thermal water deposition can only be determined in an area where thermal activity is occurring. Because the springs in the Dubois area are dead or nearly so, figures pertaining to rate of deposition are not available. The cause of thermal water deposition is chiefly due to the loss of CO_2 from the water, the result of the sudden pressure drop on reaching the surface plus the activity of algae (Twenhofel, 1950). Rate of deposition would therefore vary, being dependent on rate of flow, concentration of mineral matter, and pressure drop at any particular time. Twenhofel (1950) reports a deposition of 15 cm annually at San Vignone at San Fillippo, Sicily, 30 cm accumulated in a 4 month period. Allan and Day (1935) report that 22.2 inches per year has been the maximum rate of deposition at Mammoth Hot Springs, Yellowstone Park, with the average rate being only 8.3 inches.

TABLE 1.—*Insoluble residue analysis*

Specimen Number	Location	Specimen Weight	Residue Weight	Percent $CaCO_3$
1	B_7	10.920	.395	96.4
2	B_5	12.073	.148	98.8
3	B_6	9.823	.493	95.0
4	C_3	9.648	.420	95.7
5	C_1	12.158	.780	93.6
6	A_1	11.278	.072	99.4
7	D_8	13.818	.428	96.9
8	D_7	11.078	.293	97.4
9	D_6	11.493	.118	99.0

From a distance the travertine appears to have a mottled red, black, yellow, or gray color; however, when close the weathered surface is a light yellow-gray. When fresh the travertine is light yellow turning a creamy shade when wet. The weathered surface is always pitted and rough, and at many places it supports abundant brown-orange lichen and a black and gray fungus.

Bedding is generally apparent. Around the edges of an orifice the travertine is thin-bedded. On the sides of the domes a shingle-bedding is of common occurrence. Whether well-indurated or not, the bedding is traversed by vertical tubular passages which are the result of seepage and deposition.

Travertine Conglomerate

All of the thermal water area is underlain by a travertine conglomerate (fig. 2) which extends to the western bank of Wind River and possibly as far southeast as Dubois. Keefer (1957) cites a carbonate conglomerate in Sec. 1, T. 41 N., R. 107 W.

The conglomerate measures 40 feet in thickness in the thermal water area. The pebbles and cobbles, sometimes up to 3 feet in diameter, are consistent in type and texture with those making up the terrace deposits outside the thermal water area and consist of well to poorly rounded and angular pebbles, cobbles, and boulders of well-weathered granites, volcanics, quartzites, basalts, limestones, sandstones, and metamorphic rocks.

A great volume of water must have been

FIG. 2.—Travertine conglomerate, thermal water area.

released by the springs. As it sought the level of Wind River much of it flowed over and through the terrace deposits. This produced the cooling and retardation necessary for the deposition of any calcium carbonate left in solution, which cemented the terrace deposits into the resistive strata which today crop out on the western bank of Wind River.

In the area between Geyser Creek and Water Gulch, and removed from the thermal water deposits, travertine beds 6 inches to 2 feet thick overlie the terrace.

LOCATION

The thermal water deposists of the Dubois area occur in five separate groups, four of which are discussed in this paper (fig. 1). The fifth area, which occurs only a short distance west of Group B (fig. 1), is an area of natural bridge formation due to a combination of thermal water and erosional activity. This area was adequately discussed by Delo and Miner (1942) and more recently by the senior author's stratigraphic study of the Dubois area (1958).

For simplication, the four groups here discussed have been assigned capital letters beginning with A. Each separate spring orifice has then been assigned a number of subscript, beginning with 1; thus A_3 refers to spring orifice #3 in Group A.

Group A.—Group A consists of 5 orifices which open on the sides or near the bottom of a large travertine dome located in Sec. 23, T. 42 N., R. 108 W. The main orifice, A_1, is centrally located on top of the main dome, whereas A_2–A_6 occur on the sides.

The travertine, which was deposited by this group of springs, reached the surface which then sloped at 30–40 degrees. Deposition was initially vertical but soon spread laterally forming the terrace-like structure which exists today.

The secondary orifices, A_2–A_6, being down slope from A_1, signify that when the travertine reached a certain height, above which it became difficult to force a continual flow of water, the thermal waters sought lower outlets. The advanced erosion of A_1 supports such a dating.

Group B.—Group B consists of a series of domes which line the walls and bottom of Warm Spring Canyon from Wind River upstream to just below the Natural Bridge. Some of the domes perch on the steep walled canyon edge with no overlap of travertine, thus indicating that at the time of deposition Warm Spring Creek was either much smaller or non-existent. Younger domes exhibit a smooth profile and, when on the edge of the canyon, overlap the lower bedrock, in many places to the stream's edge.

B_1 is one of the younger domes which overlooks the entrace of Warm Spring Creek into Wind River. The side of B_1 which faces Warm Spring Creek is undercut thereby exposing a good cross-section of the deposit. Figure 3 is a diagramatic cross-section of B_1 illustrating the relation of the travertine to the underlying terrace mantle and bedrock.

The lowest part of B_1, which rests on the Phosphoria Formation, consists of an incompetent conglomerate of gneisses, schists, and granite cemented by a calcium carbonate matrix. Interbedded throughout the

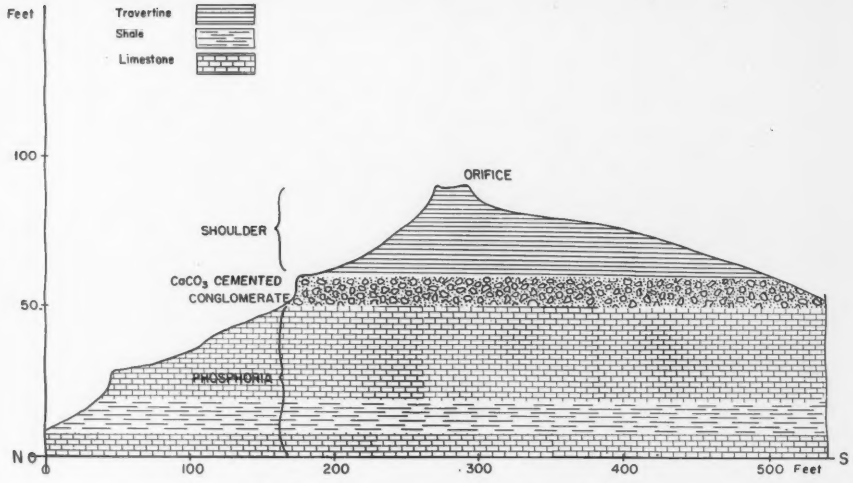

FIG. 3.—Diagrammatic cross-section of B_1.

conglomerate are cross-bedded sands and gravels from stream deposition. Above the conglomerate a 30-foot shoulder surrounds the orifice. This shoulder was no doubt at one time an orifice of considerable size; however, by the time deposition ceased the orifice had shrunk to a diameter of 25 feet. Within the small remaining orifice exists a shallow, saucer-shaped depression which is surrounded by a thin lip of travertine. This represents the final opening of the spring.

B_2 is a much older dome than B_1. Lying southwest from B_1 and south of Warm Spring Creek, B_2 is built on the side of a notch in the Tensleep hogback (fig. 4). To the north of smaller shoulder of calcareous

FIG. 4.—Northeast flank of B_2.

FIG. 5.—Cliff formed by B_4 along Warm Spring Creek.

tufa abuts B_2 and encircles the open end of the hogback. The calcareous tufa comprising B_2 is much harder than that of B_1; an example of increased induration with age.

B_3 is a much smaller deposit than either B_1 or B_2 but is of far greater interest. The north edge of B_3 forms part of the south wall of Warm Spring Canyon, the rest of the dome forming a low lying, smoothly profiled flat. Instead of a small circular orifice of limited depth, a large crack and irregular opening exist on top of the dome. With caution the geologist can safely descend into the orifice and through 50 feet of travertine to a large chamber, the roof of which shows the slump filled subsidiary orifice. Tepid water, indicating that the spring is still active, fills an indigo blue pool some 60 feet deep at the bottom of the chamber. The water is kept fresh and at constant level by circulation from its underground source which ultimately reaches Warm Spring Creek. Because of its escape into Warm Spring Creek and thence into Wind River, neither stream freezes during the winter months until reaching a spot just north of Dubois, 12 miles southeast (Ben H. Weeks, personal communication).

Upstream from the previous domes the dome designated as B_4 caps the Amsden Formation with 100 feet of travertine. B_4 is considered to be one of the oldest thermal water deposits because over one-half its original mass has been removed by the lateral erosion of Warm Spring Creek (fig. 5). This then means the spring existed while Warm Spring Creek was in its infancy. Advanced age is further signified by the degree of induration.

The thermal water deposits labeled B_5 are a series of spring remnants which line the sides and top of the north edge of Warm Spring Canyon. Erosion has destroyed original unity leaving only pedestals, chimney, and knife rocks. The travertine in this group is well-banded and exhibits numerous coarse layers.

B_6 is a young dome which occupies a small valley flat in the bottom of Warm Spring Canyon. Around the orifice exposed travertine gives a hollow sound when walked or tapped on, suggesting a very thin deposit which is out of proportion to the lateral extent. Because B_6 is so young and because of the fast rate of deposition, the channel of Warm Spring Creek was forced against the south wall of the canyon.

B_7 and B_8 occur down river from B_6 on

the north side of the canyon. B_8 is the larger; however, both springs were sufficiently large to deposit enough travertine to overlap the canyon rim and come within 5 feet of the water level. B_7 has an orifice 6 feet in diameter from which a shallow, cold spring issues.

B_9 and B_{10} are both very young thermal deposits, both in the valley flat and on the bank of Warm Spring Creek. Both have been destroyed to a great extent because of their proximity to the water level, however both still affect the stream channel, forcing it to flow around them.

Group C.—This group of thermal water deposits parallels Geyser Creek, Sec. 32, T. 42 N., R. 107 W. Geyser Creek has cut through the Tensleep hogback to join Wind River, the bottom of the cut now occupied by domes C_2, C_3, and C_4. The domes C_2–C_4 depositionally formed a three step terrace which is easily confused with the glacial terraces in the vicinity.

C_2, the farthest upstream, shows evidence of damming Geyser Creek just before it leaves the canyon cut in the Tensleep hogback. The constriction of the walls caused deposition to be nearly vertical instead of horizontal (lateral), thus the highest terrace is formed of nearly vertical thermal deposits. C_3 and C_4, lying downstream, also built terraces; however, after Geyser Creek passed through the Tensleep hogback it was able to cut laterally thus forming a wider valley the lower it flowed. This, then, allowed thermal deposition over a much greater area which reduced the height of each succeeding deposit.

C_1 is a much younger dome. Occupying the mouth of a northward flowing tributary to Geyser Creek, C_1 consists of 60 feet of travertine with an orifice 15 feet in diameter.

C_5 is by far the largest thermal water deposit of Group C. A symmetrical dome with travertine layers shingling the sides, C_5 caps the northern edge of the Tensleep hogback where Geyser Creek cuts through, forming a 35-foot escarpment facing the creek. On the northern approach tension cracks, which have been filled with mantle debris, line the sides and converge toward the orifice. The orifice is a shallow saucer-shaped depression with a 40-foot diameter.

Group D.—The thermal deposits classed as Group D parallel the strike of the outcrops. Two distinct, parallel trends are formed: D_1, D_2, D_3, D_6, D_7, and D_8 following the base of the Tensleep hogback, D_4 and D_5 at some distance from, but still parallel to, the hogback.

D_2, D_3, D_4, and D_5 are symmetrical, only 3 to 4 feet thick but of great lateral extent. The diameter of D_5 exceeds 575 feet. No orifices are found in this group.

D_1 is 25 feet thick, the lower 20 feet being a rounded base, the upper 5 feet a shoulder with a central orifice. On the inside of the orifice the travertine was undercut by the thermal waters.

D_6 and D_7, when considered as one dome, have a length of 4500 feet, width of 2400 feet, thickness of 300 feet; the largest continuous thermal water deposit in the Dubois area.

Figure 6 is a diagrammatic section through D_6. The orifice is superimposed on a broad,

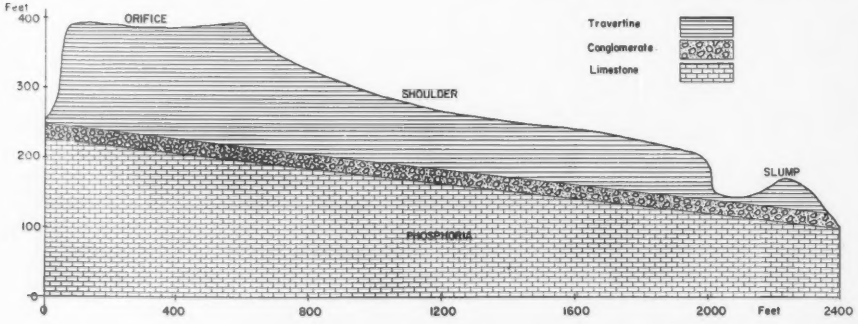

FIG. 6.—Diagrammatic cross-section of D_6.

terrace-like shoulder measuring 155 feet thick. The bulk of the calcium carbonated waters came to the surface through the orifice, spilled over and ran down dip; greatest deposition naturally took place near the orifice, the shoulder resulting from deposition of carbonate from the waters as they ran down dip.

D_7 is just as thick as D_6, however its lateral spread has been restricted due to the erosional activity of Water Gulch—the channel for the surface runoff from both springs.

D_8 represents only a remnant of a thermal water deposit which no doubt at one time extended across Water Gulch. It is possible that D_8 represents an outlier of D_7, however no supporting evidence could be found.

STRUCTURAL AND STRATIGRAPHIC RELATION

The relationship of thermal water deposition to local stratigraphy is important in determining the source of mineral enrichment. Obviously, extreme and rare structural conditions must exist before thermal waters can derive mineral content from units stratigraphically higher than those on which they outcrop. Since essentially all emanations of thermal waters are of a near vertical nature, the source of mineral content must be from stratigraphically lower beds.

The thermal deposits are located on strata ranging in age from Mississippian to Triassic, therefore pre-Pennsylvanian strata may be considered as the source of the mineral enrichment. The universal deposition of calcium carbonate strongly suggests that the Ordovician Bighorn, Devonian Darby, and the Mississippian Madison formations were the source of the carbonate.

That these formations are easily soluble is seen in the exposures along Warm Spring Canyon. There the Madison is typified by solution cavities which often extend down to the Darby. The Bighorn lacks evidence of solubility, however it cannot be overlooked as a possible source.

The presence of a thermal water deposit immediately suggests a structural weakness of some kind. The parallelism of Groups B and C perpendicular to the strike of the strata definitely suggests fault control. The thermal deposits of Group B line the sides and bottom of Warm Spring Canyon, which in some areas is over 250 feet deep. This depth and the extreme narrowness, suggest an aid in addition to running water.

The writers believe that faulting controlled the location of the deposits in the thermal water area because:

1. Warm Spring Creek could have flowed at a lower level by going around Warm Spring Mountain, or to either side, rather than through it.
2. Dips of the formations on the south side of the Canyon are very much steeper than those on the north.
3. In the NE ¼, NW ¼, Sec. 1, T. 41 N., R. 108 W., a small fault occurs in the upper part of the Madison. A displacement of 8 feet was measured.
4. The angular channel pattern of Warm Spring Creek suggests fault control.
5. Apparent displacement is noted westward of the block between Geyser Creek and Water Gulch disclosed by the Tensleep hogback offset.

Since Groups A and D parallel the strike of the strata, seepage along bedding planes is suggested: Group D seems to be related to the Permian Phosphoria-Triassic Dinwoody contact, Group A to the Pennsylvanian Amsden-Tensleep contact.

If faulting has occurred in the thermal water area resultant slippage along bedding planes would have produced paths of ascention for any subterranean fluid.

Keefer (1957) believes that the deposits coincide with the Tensleep-Phosphoria contact, especially since he dismisses fault control because "no faults are present at least west of Warm Spring Creek."

AGE OF THERMAL DEPOSITS

General agreement has not been reached as to the age of the Dubois thermal water deposits. Keefer (1957) believes the thermal activity to have been contemporaneous with the Pleistocene to Recent terrace development of the area. The authors concur that a post-Pleistocene age is immediately suggested, however, the age relationship between Warm Spring Canyon, the terraces, and the thermal deposits must be closely examined.

It is readily apparent that two distinct

periods of thermal activity occurred, the oldest predating Warm Spring Canyon, the youngest post dating the Canyon. Hence, the age of Warm Spring Canyon rather than the age of the terraces is the key to the most probable age of the thermal deposits.

General absence of any evidence of glaciation in Warm Spring Canyon suggests a pre-Pleistocene or post-Pleistocene age, the post-Pleistocene (to Recent) the most probable. Thus the travertine domes, when cut by the canyon could be as old as pre-Pleistocene. The flows which overlap the canyon rim are Pleistocene to Recent in age.

SEDIMENTARY ANALYSIS

The thermal deposits are almost wholly calcium carbonate. Samples were collected from nine locations and analyzed for insoluble residue, the results appearing in table 1.

The insoluble residue was placed in bromoform, and the light and heavy minerals separated. The light minerals consisted entirely of quartz and the clay minerals. The quartz was clear to cloudy, angular, and without crystal faces. The clay minerals, the product of weathered feldspar, no doubt originated in the Precambrian granite of the area.

The heavy minerals were separated into the magnetic and non-magnetic factions. The non-magnetic minerals were garnet, zircon, and rutile. Garnet was of the pyrope variety, crimson in color and angular. The zircon was present both in idiomorphic form, and without crystal faces, the latter variety being angular and ranging in color from colorless to light brown.

Magnetite was the most abundant of the magnetic minerals. The magnetite was angular, black and metallic with a small percentage showing alteration to hematite. The moderately magnetic garnet almandite was next in abundance, having a deep red color, irregular fracture pattern, pitted surface with iron inclusions, and a vitreous luster. Hypersthene was present, yellow-green in color and crystalline in rounded, prismatic forms. Rarely occurring was melonite, with vitreous luster, in moderately magnetic, subangular to subrounded black dodecahedral grains.

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FIGS. 1–4, SEPTEMBER, 1959

STRUCTURES IN THE STRATIFIED LATE-GLACIAL CLAYS OF WINDERMERE, ENGLAND[1]

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ABSTRACT

The succession of sediments in Windermere in the English Lake District is comprised of stratified late-glacial sands, silts and clays, post-glacial organic mud, and, at the top, surface ooze. A programme of coring was commenced in the summer of 1958 in order to study the sedimentation and sedimentary history of the late-glacial sediments. The stratification of these sediments is marked, and consists of an alternation of sand-silt and clay layers. The sand-silt layers are usually laminated and graded. It is believed that these layers were deposited from bottom-sweeping currents. The alternation of coarse and fine layers resembles varves (De Geer, 1912). It has been found that the correlation of the layers in cores is only possible over short distances; therefore, it is clear that the layers in these sediments cannot be used for dating.

Cores show that the sediments have been affected by post-depositional movements down the slopes of the lake basin. These movements range in scale from small movements affecting single layers to the oversteepening, overturning, and in some cases, the slumping of several feet of sediment.

All movements appear to have ceased before the deposition of the organic mud.

INTRODUCTION

The author recently commenced a study of the stratified late-glacial sediments found in cores collected from Windermere, largest lake of the English Lake District. Windermere is a lake of glacial origin, having been formed during the Quaternary glaciation by a combination of over-deepening and morainal damming of the exit. The lake has two distinct basins separated by a line of low islands and shallows. The north basin, shown in figure 1, is about 4 miles long with a maximum width of nearly one mile and a maximum depth of 220 feet (68 metres). The longer and narrower south basin is 6 miles long, with a maximum width of one-half mile and a maximum depth of 138 feet (42 metres).

This paper describes some of the structures found in the stratified clays collected from the north basin during the preliminary coring programme. The cores were collected by means of a Mackereth Core Sampler, details of which have already been given (Mackereth, 1958; Smith, 1959).

The greater part of the investigation was carried out at The Ferry House, the Windermere Laboratory of the Freshwater Biological Association. Earlier workers at this laboratory have made detailed studies of the post-glacial sediments of Windermere and the present study is, in a way, a continuation of their work.

[1] Manuscript received December 26, 1958.

DESCRIPTION OF THE DEPOSITS

Three main types of deposit exist in both basins. These are: (1) surface ooze, (2) organic mud, and (3) stratified clay.

(1) The surface ooze occurs everywhere except in very shallow water (less than 6 feet). Its thickness ranges from about 1 inch, in fairly shallow water, to about 10 inches at the greatest depths of the lake.

(2) The organic mud underlies the surface ooze. It contains about 25 per cent (dry weight) organic matter. Detailed descriptions of the organic material have been given by Pennington (1943, 1947). The mud ranges in thickness from about 3 feet, in shallow water, to about 16 feet, in the deeper parts of the north basin. The greatest thickness of organic mud so far found occurs near the northern end of the north basin, where 23 feet of organic mud were encountered without reaching the lower limit.

(3) Below the organic mud there is usually a narrow transition zone of mixed mud and clay. This zone is underlain by the stratified clay. The maximum thickness of this clay is not known. Near the lake shore the corer has penetrated the clay and has reached pebbly sands. Here the thickness

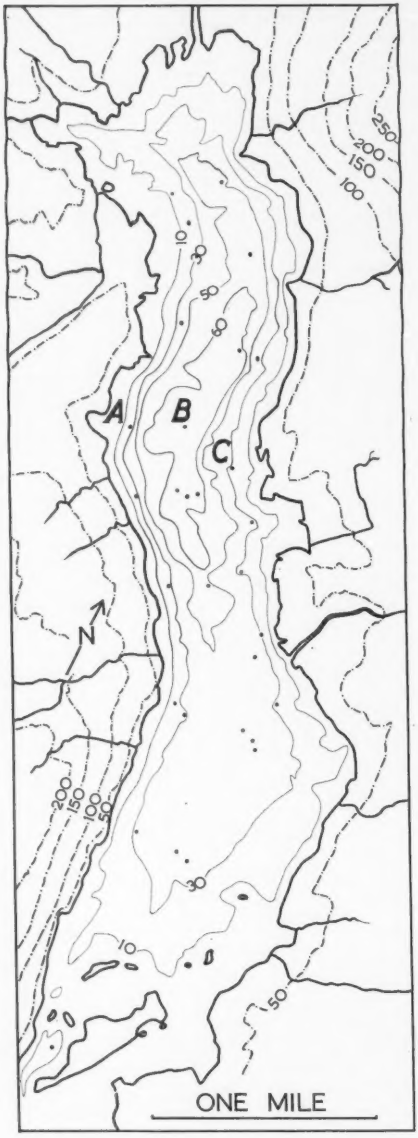

FIG. 1.—Map of the north basin of Windermere showing the positions of the core stations which are marked by dots. A, B, and C mark the position of three core stations which are referred to in the text. All heights and depths in metres.

of the clay may be as little as 1 foot 6 inches. Farther from the shore the clay thickens and is underlain by fine sand and silt (0.15 mm–0.05 mm diameter). The passage up from sand to clay is a gradual one. The upper part of the sand is stratified, the sand alternating with silts and clays. Fine sand layers often persist for a considerable distance into the stratified clay. Many cores collected from near the lake shore show a silty layer with a high organic content near the top of the stratified clay.

It is thought that the stratified clay is of late-glacial age and was deposited in fresh water. Although rock fragments are an important constituent of sands and coarser factions, quartz appears to be the predominant mineral constituent of the silts and clays.

STRUCTURES FOUND IN THE STRATIFIED CLAY

The layering in the stratified clays is a striking feature. The stratification appears as an alternation of light and dark material. The difference in colour does not appear to be due to any mineralogical variation; but, rather, it is due to variations of grain size. The light layers are composed of very fine sand or silt (0.125 mm—0.004 mm diameter) with some material of clay dimensions, whereas the dark layers are composed entirely of particles of clay dimensions. The colour difference appears also to be dependent upon the water content. The layering only becomes apparent on the partial drying out of cores; cores which have been dried out completely do not show any noticeable layering.

Of the 34 cores which have been collected so far, 29 are from the north basin, and 5 are from the south basin. Twenty-four of the 29 cores taken from the north basin contained stratified clay. While all the cores containing stratified clay showed horizontal or nearly horizontal layers near the passage into the organic mud, only 7 cores showed horizontal laminations for the whole length of stratified clay. Five cores showed undisturbed inclined layers, this inclination varying between 5° and 35° for part of their lengths, though the topmost layers were horizontal or nearly so. The higher angles are usually found in the lower part of the cores. The remainder of the cores showed disturbed layering for some part of their

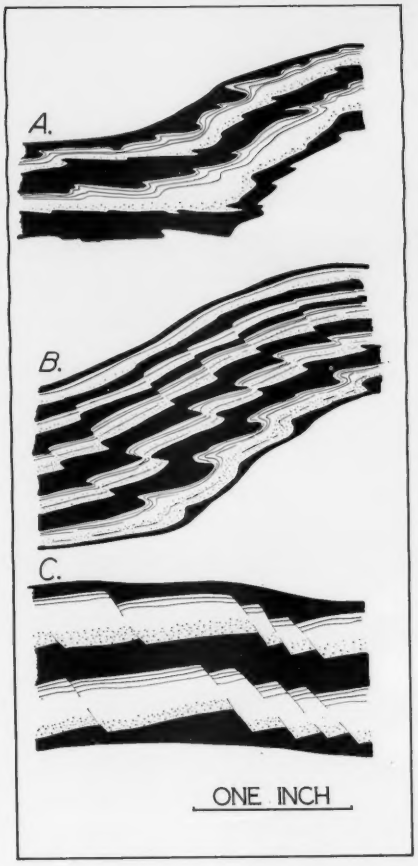

FIG. 2.—Minor structures in the stratified clay. A—"Crinkle marks"; B—Reversed "faults"; C—Normal "faults." Descriptions in the text.

lengths. This disturbed layering is usually associated with inclined layering. It is quite clear that a considerable amount of post-depositional movement must have occurred prior to the deposition of the overlying organic mud.

The undisturbed clays.—The thickness of the layers ranges from approximately 3 inches, for light coloured layers, towards the bottom of the stratified clay, to 1/32 inch. for light layers occurring towards the top. Generally speaking, there is a gradual reduction in the thickness of the light layers from bottom to top of the stratified clays, though the thin 1/32 inch layers occur throughout many of the cores. The dark layers also show a general thinning but it is less marked than in the case of the light layers. In thickness, the dark layers do not exceed 1 inch in the lower part of the stratified clays, yet, they rarely are less than 1/10 inch thick.

All but the very thinnest light layers are laminated. Although the laminae are very thin, each lamina shows a vertical grading which is made apparent by a change in colour. Clay size particles occur throughout each lamina. Any light layer composed of laminations also shows a form of grading, the lowest lamina containing the coarsest grains in the layer, and successive laminae showing a reduction in the size of the coarsest particles present in them. The thickness of most complete laminae is about 1/32 inch. There are a few unlaminated layers up to ⅛ inch thick. These layers are always graded.

All layers show sharp lower surfaces; the upper surfaces are transitional into clay.

The disturbed clay.—Most of the core samples from the stratified clay show disturbed layers which always are associated with inclined layers. The irregularities may be of a minor nature, where each layer reacts individually to new conditions, or of a major nature, where a group of layers have been affected and have reacted as a single unit.

Minor structures take three forms. The first form appears as minor overfolds in a light layer (see fig. 2A). The upper and lower surfaces of the layers are affected by the folding, and the structure of the fold indicates a movement down a slope. It is suggested that these structures, which resemble very closely the "crinkle marks" described by Williams and Prentice (1957) from the Ludlovian rocks of the Welsh borderland, result from flowage of the plastic clays while under pressure from above and behind, that is, from the up-slope direction.

The second structure generally occurs in slightly thicker layers and is closely associated with crinkle marks. It takes the form of small thrust planes cutting the light layers (see fig. 2B). The small thrust planes generally start near sharp folds in an undulating "crinkled" layer. The thrusts

rapidly die out upwards and usually only affect two or three layers. The layers affected by thrusting do not show any small-scale folds.

FIG. 3.—Major structures in stratified clay.

The third minor structure has the appearance of small normal faults (see fig. 2C). The small faults do not cross more than one layer, though successive beds may be affected in the same way. This last structure is thought to have resulted from tensional forces, the tension having been caused by the sliding of sediments further down the slope resulting in a pull on the up-slope sediments. This structure is similar to the "pull-apart" structure described by Natland and Kuenen (1951).

The major structures show a considerable variation in size and may affect several feet of sediments. The structures range from the over-steepening of the layers to the complete collapse of the original bedding. In some cores over-steepening is accompanied by overturning. If the over-turning is on a large scale it is topped by a slide plane succeeded by layers which are right-side up (fig. 3). The lowest laminated layers above the slide plane usually show either crinkle bedding or small thrusts, or both. The large scale structures are thought to result from over-loading in the up-slope areas leading to gravitational sliding. These structures occur in the fine grained upper parts of the stratified clay. The sand and silt layers in the lower part of the clay, when affected by gravitational sliding, show minor faults which may cross several layers.

In the extreme form of post-depositional movement complete disruption of the layering occurs. Whole sections of the core are composed of slumped masses consisting of slump balls, showing contorted laminations, surrounded by a matrix of structureless clay. Such slumped masses succeed inclined layers which are often over-steepened or over-turned. At the base of the slump there is a slide plane which truncates the over-turned layers. The slump is usually succeeded by inclined layers. Many of those layers nearest the slump are irregular and show the minor structures described above. Some slumps show a graded layer of unsorted sand and silt which passes upward into structureless clay with slump-balls as shown in figure 4. In this core the graded layer appears again higher in the slump mass where it appears to be overturned, suggesting that the slump mass has undergone later movement.

FIG. 4.—Part of core showing slumped sediments. Laminated fragments in a structureless clay matrix succeeded by thin layers of silt and clay. About two-thirds actual size.

Besides the planes of movement described above as minor structures, some cores which include a considerable amount of the stratified clay show larger reversed "faults" in their lower, silty, layers. Cores which show the "faults" also show either over-steepened layers or slumped masses in the higher, fine grained part of the core. It is suggested that these "faults" are the expression of readjustment in the silts and were probably formed at the time of the over-steepening or slumping of the higher layers.

CORRELATION OF CORES

A correlation of cores has been attempted, but results so far have shown that complete correlation of the layers over any large part of the lake is not possible. Some cores collected at distances up to 100 feet from each other have been correlated; although, even at these distances, discrepancies have been noted. Over greater distances correlation has not been possible. It is hoped that a detailed picture of the variations will be built up as coring proceeds. Only one instance of correlation over a considerable distance has been noted so far. This occurs between cores taken at stations 8 and 10, which are nearly 2,000 feet apart (marked A and C respectively in figure 1). Station 8 is on the western slope of the lake basin, depth 29 metres; station 10 is on the eastern slope at 28 metres. The core collected at station 9 (B in figure 1), half-way between 8 and 10 in the lake centre at 63 metres, did not reach the stratified clay, and is entirely composed of organic mud.

Even in this instance only part of the cores can be correlated closely. The lower portions of both cores are composed of quite thick (1 to 2 inch) silty layers alternating with thin clay layers ($\frac{1}{2}$ to 1 inch thick). Because of the high water content of the silts, these layers collapsed on being removed from the core tube and thus no precise correlation of them was possible. The succeeding finer grained layers can be correlated very precisely. The total length over which correlation is possible is $17\frac{1}{2}$ inches of core 8 and $16\frac{1}{4}$ inches of core 10. Thirty-five obvious alternations of light and dark material exist in these lengths, the light layers showing a gradual reduction in thickness

from ½ inch at the lower part to about 1/20 inch towards the top. Detailed correlations of layers can be made between the cores; features characterising certain layers are common to both cores, and many layers are comprised of the same number of laminae.

Overlying the parts of the cores that can be correlated both cores show 8 inches of clay with thin laminations (about 20 light and dark layers in an inch). However, slight variations exist between the cores and no definite correlation is possible. Above this portion of the cores similarity does not exist. Core 8 shows 10 inches of unlaminated clay showing slight colour variations succeeded by 4 inches of clay with fine laminations (each lamination is about 1/20 inch thick). Above, there is a transitional zone 17 inches thick composed of a mixture of grey clay and organic mud. This is overlain by organic mud. Core 10, in contrast, passes from the clay which can be correlated, into 6 inches of grey clay exhibiting only a very indistinct layering. This is succeeded by 27½ inches of laminated clay. The laminations are thin, the light, coarser, layers averaging about 1/20 inch thickness, while the darker layers are between 1/15 and 1/10 inch thick. Over 170 light layers can be discerned. A ¼ inch thick silt layer, unlaminated, but graded, occurs within this laminated clay. The laminated clay is succeeded by a transition zone 3 inches thick.

There is no suggestion of a sudden break which would indicate that slumping could have caused the discrepancy above the identical parts of the two cores.

DISCUSSION AND CONCLUSIONS

Glacial and late-glacial sediments which are banded or laminated and show an alternation of silt and clay layers are usually called *varved* deposits. It is generally accepted that each light and dark layer couple is deposited in the course of a climatic cycle, that cycle being a year. The term has come to be associated with late-glacial sedimentation, the light coloured sand and silt layers being deposited during the spring and summer thaw when turbulent conditions prevailed, whereas, the fine grained dark layers represent the deposits of the stagnant conditions which existed during the winter. De Geer (1912) believed that this was the case and that varves could be used for dating and for correlation, even between continents (De Geer, 1921). It is tempting to regard the repetition of light and dark layers in the stratified clays of Windermere as an annual layering that may, as such, be used for dating. However, there is no clear indication that the light and dark layers are the result of annual climatic changes. It is possible that the alternation of light and dark layers may reflect local conditions. Hansen (1940) in his description of the late-glacial sediments of Denmark and southern Sweden suggests that the deposit of one year may be made up of many light and dark layers. The marked dissimilarities of cores collected from Windermere make it impossible, at present, to use the layering for dating.

The cause of variations between cores is one of the problems being studied. It is clear that the sediments did not enter the lake-basin at one point, rather the sediments entered the lake-basin at many points and were deposited as large submerged delta-like features, the sediment being derived from the meltwaters of the retreating ice.

It is thought that the sediments spread out from the mouths of streams as bottom-sweeping clouds of suspended sediment; the turbidity currents of Natland and Kuenen (1951). (The term "current" is unfortunate since it suggests restricted transverse dimensions and this is not thought to have been the case here.) Deposition from such moving clouds would explain the graded bedding found in the light, coarser, layers of the stratified clay. Besides this vertical grading the moving clouds would lead to horizontal grading. The laminations found in the light layers possibly reflect deposition from several clouds: the "grading" of the laminae may have resulted from the ability of a denser cloud carrying coarser sediment to pass under a cloud carrying finer material.

The direction of travel of such clouds would be controlled by the contours of the lake floor. While the cause of any flow of sediment may have been common to all parts of the lake the final amount of sediment and the number of layers found at any point would depend upon the local configuration of the lake floor.

Variations in the clays could be the result

of slumping of sediments. Slumping would remove the record of deposition from one point, confuse the record at another, and, in the event of the slump being completely mixed with water, result in many layers being reduced to a single graded layer at a third point. Slumping is thought to have been caused in two ways: either by the sediments moving forward under their own weight during or soon after deposition; or by movement at a later stage when the lake level was falling. There are definite indications of lowering of the lake level. Under such conditions there would be excessive loading of the sediments around the lake which would cause instability and slumping.

The reduction in the thickness of the light coloured layers in the stratified clay may reflect the retreat of the ice-cap. De Geer (1912) suggested this as the cause of the thinning of Scandinavian varves. However, it seems more likely that the thin light-coloured layers are related to the lowering of the lake level and represent redeposition of sediments as a result of erosion.

From this account if may be seen that a considerable variety of structures exists in the late-glacial clays of Windermere and that with an intensive programme of coring it should be possible to determine the sedimentary history of the lake.

ACKNOWLEDGMENTS

The author would like to thank his wife, for her valuable assistance both in the collection of the cores and in the laboratory; and H. C. Gilson, the Director of the Windermere Laboratory of the Freshwater Biological Association, for the use of equipment belonging to the Association. Valuable assistance during coring operations was given by A. E. Ramsbottom of the Windermere Laboratory. The present study has been partly supported by a research grant from the Central Research Fund of the University of London.

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FIG. 1, SEPTEMBER, 1959

THE RELATION BETWEEN IRON AND ORGANIC MATTER IN SEDIMENTS[1]

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ABSTRACT

Use of the densitographic method shows that the iron in sediments forms complexes, chiefly of a physical nature, with organic matter. The complex which is formed contains, in 80 percent of the cases, about three parts by weight of organic matter and one part of iron oxide. There are notable exceptions. Beach sand does not contain any appreciable amount of organic matter, and in certain "hardpans" the iron content is much higher than that of organic matter. In other sediments the organic content is greatly in excess of the iron. It is suggested that other metals, for example uranium in black shales, may form similar complexes.

MATERIALS AND METHODS

The materials used are listed in a previous paper in this journal (Baas Becking and Moore, 1959). They comprise nearly one hundred samples of estuarine, marine and freshwater sediments and a few samples of sedimentary rocks. These samples were submitted to densitometric analysis. All samples were examined microscopically. If magnetite, pyrite or marcasite were detected, the iron was extracted by treatment with 60% perchloric acid. In other cases 2N hydrochloric acid sufficed. The iron was then determined colorimetrically in a Lange photometer or a Spekker adsorptimeter, using the thiocyanate method.

Organic matter was determined either in the wet way (Piper, 1947, p. 214), by digestion in 20% hydrogen peroxide or, in the majority of cases, by ignition, subtracting 6% of the weight of clay minerals present. The figures for organic matter are less reliable than those for iron.

IRON CONTENT OF THE SEDIMENTS

The iron occurs in the samples as hydroxide; there may be phosphate, magnetite, siderite, and sulphide in addition. In this paper it is expressed as FeO(OH). In shales (Clarke, 1916) the ratio ferric/ferrous iron is about 1.6, in sandstone it is around 3.3, and no ferrous iron is reported from limestones. In foetid limestones nearly all the iron should be in the ferrous state, so that the above results may be ascribed to insufficient sampling. Figure 1 shows the iron content of the materials used by the authors. The distribution is given in table 1; it appears to be logarithmic. The mean is almost identical with that obtained by Rochford (1952) for the average composition of estuarine muds. Clark's results are shown in table 2. The sediments shown in table 1 comprise 60% shales, 24% sandstones, and 16% limestones (estimation from a previous paper by the authors (Baas Becking and Moore, 1959)). Using Clarke's figures for iron hydroxide content we obtain an average of 5.03% FeO(OH), which is in close agreement with our own figure of 4.71%. The influence of sulphate-reduction on a water depends primarily upon its sulphate content. For seawater $SO_4^{=} = 6.2 \times 10^{-2}$N, and excess of hydrogen sulphide will be produced at complete sulphate-reduction if Fe^{++} is less than 6.2×10^{-2}N, that is, less than 0.17% Fe^{++} or 0.27% FeO(OH). As nearly all this iron is available only in the mud surface only a fraction of it will be able to bind hydrogen sulphide. Moreover, "calamitous" sulphate-reduction will take place when a sulphate-rich water becomes isolated. The depth of the isolated water masses is variable. In cases observed by the authors in estuaries it may vary from 1 to 10 metres. Iron oxide content of the mud lower than 2% could lead to the formation of "foetid black mud" with a complete exhaustion of the water-

[1] Published with permission of the Director Bureau of Mineral Resources, Geology and Geophysics. Manuscript received August 12, 1958.
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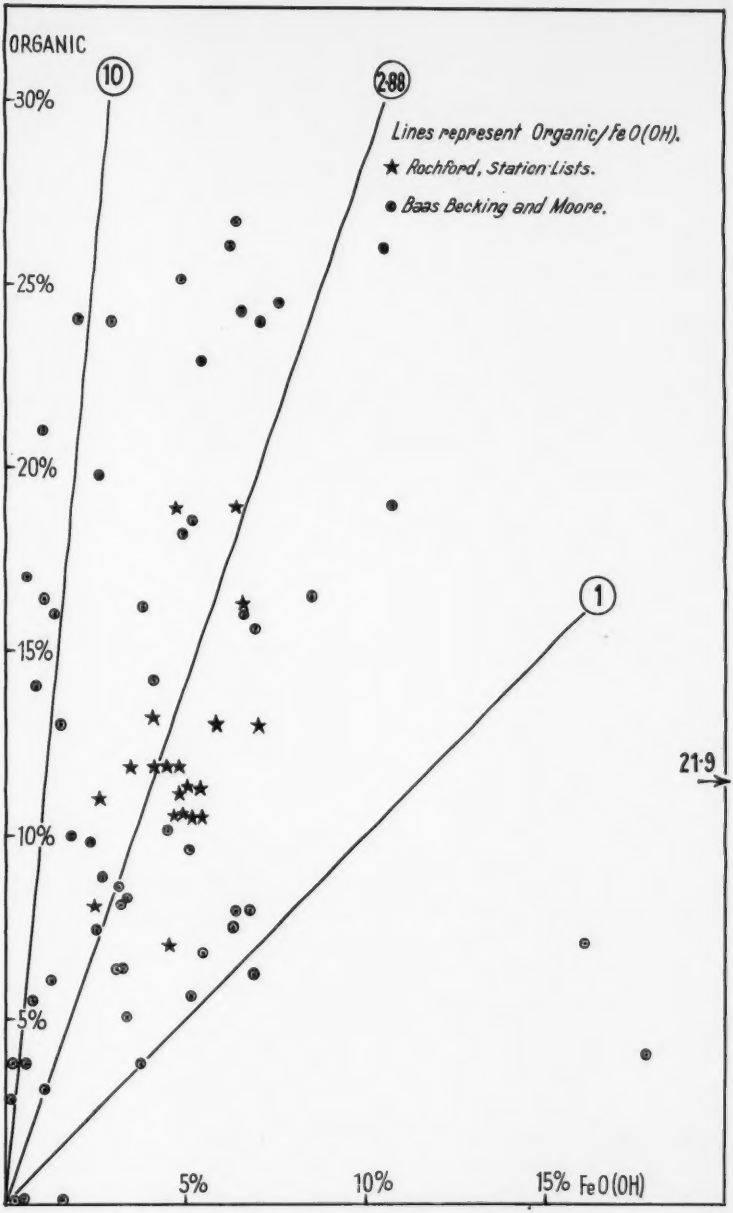

FIG. 1.—Relation of organic matter to iron hydroxide in muds.

TABLE 1.—*Iron content of sediments*

% FeO(OH)	0	2	4	6	8	10	12	14	16	18	20	22
% variant	29	22	20	21	1	4	0	0	3	0	1	

Mean value = 4.71% FeO(OH).

mass in oxygen. About 30% of the determined iron values are of this order, two-thirds of them from marine and estuarine muds. On the other hand, very high iron values were obtained from a hardpan in coral rubble from Isle Charron, New Caledonia, and in a gravel layer from the bottom of a 10 foot core from the freshwater Lake George, New South Wales.

ORGANIC CONTENT OF THE SEDIMENTS

The nature of the organic matter in the sediments may be ascertained from microchemical reactions. In recent sediments this material may be pectic, chitinous, cellulosic or lignic. In fossil sediments dehydration may give rise to "caramels" and phytomelanins rich in carbon; dehydration plus de-oxygeneration will lead to the lignite-coal-anthracite series, whereas decarboxylation might cause the accumulation of hydrocarbons.

Pectic materials are found in estuarine muds in all stages of decomposition. The authors believe that the pectinoids are the chief fuel for microbiological reactions in estuarine mud. *Chitinous* materials are abundant and are also decomposed by bacteria, aerobically as well as anaerobically, although the turnover is much slower than in the pectinoids. *Cellulose*, present chiefly in plant remains, and to a lesser extent in Tunicates, is very slowly decomposed anaerobically in seawater. The slowest turnover is shown by the *lignic* materials. Decomposition, even in an alkaline marine environment, takes a very long time, probably at least a few years. In artificial muds described in a previous paper (1959) the rates of turnover of the materials mentioned above were clearly apparent. The organic materials are even more diversified than are the iron compounds, and generalizations are, therefore, dangerous.

TABLE 2.—*Iron in sediments, after Clarke, 1916*

	Shales	Sand-stones	Lime-stones
Number of specimens	78	253	345
% Fe_2O_3	4.03	1.08	0.54
% FeO	2.46	0.30	nil
Total iron (calculated as FeO(OH))	7.65	1.57	0.59

In the samples studied by us the organic matter varied between zero and 26 percent. Much higher values are recorded in the literature, especially from freshwater sediments, and much lower values are recorded for the marine environment outside estuaries. According to Rochford (1952) the organic content of estuarine mud ranges from 8.5–19% with an average of 11.76% which is very close to the average of our results. The distribution described in table 2 still shows considerable irregularity, in spite of the large frequency classes. Analytical errors play a subordinate role here, as shown by trial determinations using different methods. It must be that heterogeneity of the organic matter is the chief cause of this irregularity.

RELATION BETWEEN IRON AND ORGANIC MATTER IN MUDS

Since the densitograms described in a previous paper (1959) showed usually the absence of free iron compounds, it was assumed that iron enters into combination with other components of sediments. There is ample evidence in the literature to support this view. As far back as 1897 Spring has shown that "the organic matter of natural waters is incompatible with iron, the two substances separating out as a flocculent precipitate." According to Spring one part of colloidal ferric oxide (presumably FeO(OH)) will remove ten parts of humus from solution. The use of iron salts or alum in filtration plants is an application of this principle. Although Clarke (1916) claims that minerals in finely divided form are carried down mechanically in this coagula-

TABLE 3.—*Organic content of sediments*

% organic matter	0		4		8		12		16		20		24		28
% variants		13		22		20		7		18		15		5	

Mean = 12.30%

TABLE 4.—*Ratio between organic matter and iron*

Ratio organic/FeO(OH)	0	1	2	3	4	5	6	7	8	9	10
% variants	8	19	27	14	11	3	3	1	0	0	10

Mean at 2.88 excluding values greater than 10 and at 3.57 including these values.

tion, there is much evidence to challenge this opinion, as the iron might well be adsorbed by both quartz and clay minerals.

Early in our work it was observed that the average density of a sediment decreases with increasing iron content. The highest density determined for normal sediments was 2.82. From this the following relations could be expressed by:

$$\text{density} = 2.82 - 0.065\ (\text{FeO(OH)})\%$$
$$\text{or density} = 2.82 - 0.022\ \text{organic}\ \%$$
$$\text{organic}\ \% = 2.88\ \text{FeO(OH)}\ \%$$

These relations are no more than tendencies. These tendencies no longer exist if there is a great preponderance of either component. However, they are explicable if we assume that the iron enters into combination (either physical or chemical) with the organic matter. In the case of the pectinoids such a combination may be partly chelate (Baas Becking and Mackay, 1955).

Figure 1 shows the relation between FeO(OH) content and organic content of 77 sediment samples; twenty of these results are taken from Rochford (1952). Table 3 shows the distribution of the relation between iron and organic matter. Exceptions exist in the region of a very low value of the quotient in iron-rich tills and in beach sand without organic matter. High values for the quotient, approaching those cited by Spring (1897), are found in Lake MacQuarie (estuarine) muds, in Cawley Bay (also estuarine) and in a deep-sea mud from the Tasman Sea. Coal represents, of course, an extreme case. The data of Rochford (1952) also indicate that the quotient between organic matter and iron oxides cannot be far from 3.0.

If there exists an organo-iron complex its specific gravity can now be approximated. Allowing a range between 2.5 and 3.5 parts of organic matter to one part of FeO(OH), the density range for iron-pectin complexes is 2.19 to 2.02, and for iron-cellulose complexes it is 2.24 to 2.08. Though sufficient analytical figures are not yet available, it appears that most of the iron is usually present in the sediments in the density fraction 2.0 to 2.3.

TABLE 5.—*Organic matter and iron in Pt. Hacking River*

Location	% organic	% FeO(OH)	% hydrated silica
Southwest Arm	62.5	30.6	6.9
Audley	71.8	22.2	6.0

Average quotient: organic/FeO(OH):2.56.

In two sediments from Port Hacking River the density fraction 1.87 to 2.37 was isolated. Table 4 shows the results of analyses of this fraction. In an artificial mud containing calcium carbonate, quartz and pectin, the fraction with density less than 2.37 was isolated after four months incubation. This fraction contained no quartz or carbonate, only organic matter and iron oxide, in the proportion 2.15:1.

DISCUSSION

Use of the densitographic method shows that most of the iron in sediments exists in complexes, chiefly with organic matter. Experience with artificial muds has shown that sulphate-reduction promotes the formation of such compound structures. Experiments with other metals, with and without iron, are under way. Though it is too early to re-

port results, it may be that uranium also gathers into an organo-complex with pectin derivatives; the proportion 500 organic to 1 uranium has been observed in uraniferous black shales in the United States (Tourtelot, 1956). It may be that a further study of this matter will lead to interesting and unexpected results.

The authors are indebted to Professor G. W. Leeper (Melbourne University) for criticism and to Mr. D. Izard for much of the analytical work.

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FIGS. 1–5, SEPTEMBER, 1959

RAPID MACROSCOPIC FABRIC STUDIES IN DRILL-CORES AND HAND SPECIMENS OF TILL AND TILLITE[1]

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ABSTRACT

Oriented till cores and hand specimens may be used for rapid macroscopic fabric studies by measuring the apparent long axes of particles on dissected or broken till surfaces. Preparation and measurements require only up to half an hour per sample. Maxima in these fabric diagrams are not as pronounced as in those obtained by more precise techniques, and the transverse maxima tend to be exaggerated. In spite of these inaccuracies, the fabric maxima are found to be in close agreement with the known ice flow direction in 16 out of the 21 till samples measured. The described procedure is suitable also for rapid fabric studies in tillites.

INTRODUCTION

Studies of alignment of pebbles in till have become an indispensable tool for deciphering movements of Pleistocene glaciers, particularly in areas with several till layers. Various techniques have been applied for investigations of till fabric. During the first twenty years after Richter published his method in 1932, most measurements were done in the field. Requirements for detailed or specific investigations of tills and other clastic sediments gradually led to development of laboratory techniques of measurement, such as Krumbein's (1939), and Harrison's (1957a) for macrofabrics, Dapples' and Rominger's (1945, p. 249–250) and Seifert's (1954) for microfabrics. Some of the laboratory methods made it possible also to study drill cores and oriented hand specimens taken from places inconvenient for field measurements. Unfortunately, all the published laboratory techniques are time consuming. Thus, Harrison's method—including preparation of sample, measuring pebbles, statistical evaluation of results, and construction of fabric diagram—requires nearly six hours per sample (Harrison, 1957a, p. 104). Microfabric studies, requiring also impregnation of till and careful preparation of sections (Sitler and Chapman, 1955), may take even a longer time.

A considerable length of time needed for detailed till fabric investigations makes them inconvenient for large-scale regional or stratigraphic studies. As local deviations of glacial movements or postdepositional fabric changes are common (Harrison, 1957b, p. 279–284), only a relatively high number of well-spread locations of measurements or sampling may bring out regional trends of glacial movements. Several tens of samples may have to be studied; and if each requires approximately six hours, many geologists prefer not to use them at all. Thus in order to encourage more till fabric investigations, less time-consuming laboratory methods have to be developed. Admittedly, some of the field techniques not discussed in this paper are relatively rapid (the original Richter (1932) method and the even faster MacClintock (1959) method require not more than one hour per section). Nevertheless, some geologists still consider their field mapping season too short for their spending a few hours at till fabric measurements in field. If they can collect oriented till samples of the size of hand specimens, they may do the fabric measurements later, after the field season, or during rainy days. Rapid laboratory methods for till fabric studies may also promote use of oriented drill cores and tillites for detailed stratigraphic investigations.

TILL CORES

When drill cores of compact till from the St. Lawrence Seaway area became available for further studies, J. N. Harris suggested cutting them in horizontal slices for measuring orientation of elongated pebbles. Such a macrofabric investigation is essentially similar to the microfabric method of Dap-

[1] Manuscript received March 9, 1959.

ples and Rominger (1945, p. 249). The difference is that no time-consuming impregnation of till and no thin sections are required. Cutting of cores with a diamond saw and subsequent washing and brushing of the dissected surfaces take up to ten minutes; measurements of 100 elongated pebbles take fifteen to twenty minutes (see Appendix for description of the procedure). Thus a two dimensional macroscopic till fabric analysis may be accomplished in half an hour, if oriented till samples are supplied.

EVALUATION OF RESULTS

This method also has some disadvantages. First of all, the elongated shapes of pebbles or granules, visible on the horizontal sections, are not parallel to their long axes in many cases. If the long axis of a blade-shaped particle is inclined and the intermediate axis is horizontal, a horizontal cut through such a particle will be elongated parallel to the intermediate axis. Such particles will tend to increase the transverse maximum. This was found to be true, when studying core no. 1: the transverse maximum dominated in four out of eight cuts (fig. 2-D, E, F and H). Subsequent measurements of true long axes of granules and pebbles in one of the slices produced a clear parallel maximum (fig. 2-H′) which was also in agreement with the glacial movement known from previous work by MacClintock (1958, p. 9). While measuring the true long axes of pebbles, the writer also noticed their imbrication towards the northeast. Very good fabric diagrams were obtained from the second till core, taken near the first one (fig. 3), because the horizontal axes of pebbles were nearly horizontal.

Secondly, if both the long and the intermediate axes of particles are inclined, the horizontal cuts through them may appear elongated in any direction, other than the parallel or the transverse. Such pebbles will reduce or spread out the peaks in fabric diagrams or, if inclination of the blade-shaped particles follows a definite pattern, the maxima in fabric diagrams will be shifted away from the true parallel or the transverse trends. In half of the studied cuts, the till fabric maxima were indeed less pronounced than in measurements of the true long axes.

HAND SPECIMENS OF TILL

For comparison, orientation of the visible elongation of pebbles was measured also in six hand specimens of till. Their more or less horizontal surface was merely brushed (not cut); it was washed with water spray and then dried. As pebbles and granules were only slightly exposed, it was not possible to teli whether their apparent elongation meant the long or the intermediate axis. Nevertheless, in all six cases the main fabric maxima were parallel to the true long axes of pebbles or parallel with striae right underneath the examined till (see four of them in fig. 4-A, B, C, E). The transverse and parallel maxima were of nearly equal strength in two cases out of six (fig. 4-B and E), with the parallel maximum ($\pm 20°$) being slightly stronger. As no cutting was required, till fabric measurements of each sample were completed in 20 minutes.

CONCLUSION

In spite of certain discrepancies between the maxima of true long axes and the visible apparent elongations on both cut and uncut till surfaces, most of the rapid determinations were in relatively good agreement with the known direction of glacial movement (16 out of the total of 21 samples measured). These results were even better than the results expected theoretically.

ALTERNATE PROCEDURE

For more reliable results a slightly longer procedure, which requires approximately one hour, may be used. This is the usual field method of measuring long axes of all elongated particles, as suggested by Richter (1932), or measuring the flat-lying, blade-shaped particles (Dreimanis and Reavely, 1953). The till sample has to be wet, to facilitate removal of matrix. If a drill-core is taken, it is split horizontally, avoiding large pebbles. A pointed knife and a brush are used for exposing pebbles and granules. All elongated particles longer than 2 or 3 mm are measured. Results of such fabric measurements are shown in figures 2-H′ and 3-C″.

TILLITE

To the knowledge of the author, no macro-

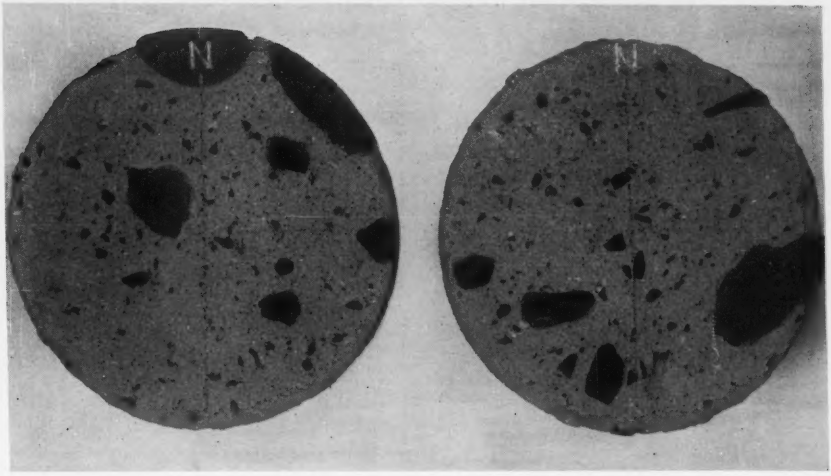

FIG. 1.—Two sections through till cores, with pebbles stained black. On the left—section St.L.I-E (fabric diagram E in fig. 2), on the right—section St.L.II-E (fabric diagram E in fig. 3). Diameter of cores $3\frac{3}{4}$ inches.

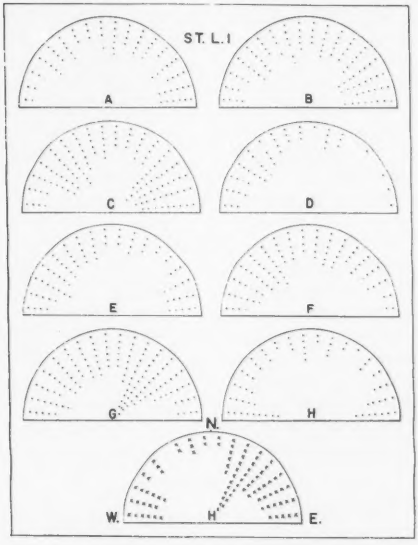

FIG. 2.—Till fabric diagrams from an oriented 1.3 feet long vertical drill core of the lower till, east end of the Barnhart island, St. Lawrence river.

Each dot represents one measurement of apparent long axis of pebble, as visible on the horizontally dissected till surface (see fig. 1); each cross—measurement of true long axis of an elongated pebble in till.

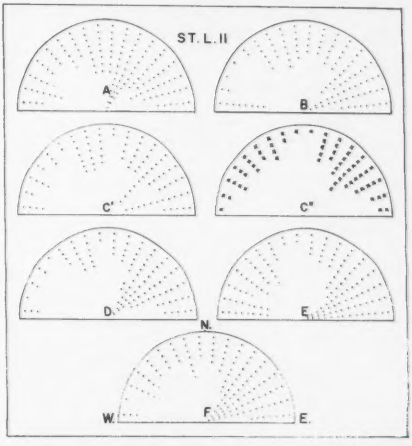

FIG. 3.—Till fabric diagrams from the same location as fig. 2; a different one foot long core at the same elevation.

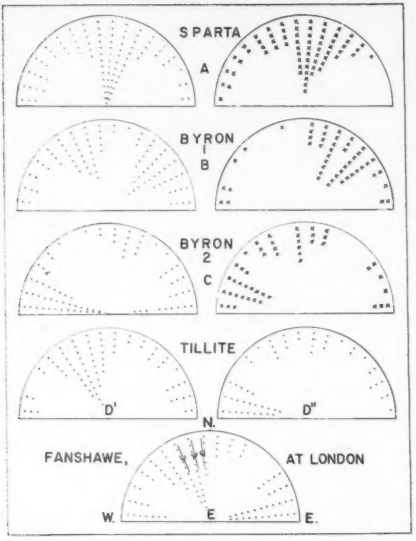

FIG. 4.—Fabric diagrams of till and tillite hand specimens from the following locations:

A—lower till, west bank of Catfish creek, 1¾ miles N.E. of Sparta, Ontario, 15 feet above river; hand specimen 3.5 by 9 inches.

B—lower till, south bank of Thames river, ¼ mile N.W. of Byron, Ontario, 14 feet above river; hand specimen 3 by 5 inches.

C—upper till, the same location as C, 45 feet above river; hand specimen 3.5 by 4.5 inches.

D—tillite, highway 11 cut 9 miles south of Latchford, Ontario; D'—hand specimen 2.5 by 3 inches, D"—1.5 by 5 inches.

E—base of middle till, south bank of Thames river, half a mile below Fanshawe dam near London, Ontario; 12 feet above river; the ice flow direction is indicated also by striae (shown by conventional symbols) on the underlying boulder pavement; hand specimen 5 by 7 inches. See fig. 2 for explanation of dots and crosses.

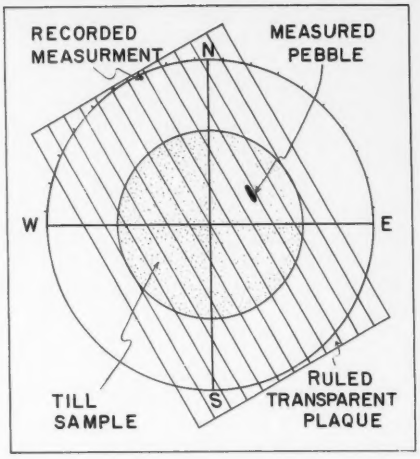

FIG. 5.—Equipment and procedure for measuring orientation of macroscopic elongated particles in till (explanation in Appendix).

fabric studies have been attempted in Canadian Precambrian tillites. If their approximate stratification has been deciphered, hand specimens of tillite should be selected with their largest surface parallel to the bedding. Elongation of the visible grains longer than 2 mm might be measured on this surface by a method similar to the one suggested in the Appendix of this paper.

This procedure was tested on two flat slabs of tillite taken from an outcrop 9 miles south of Latchford in Central Ontario. Their surfaces had been polished by Pleistocene glaciers, and all the larger clastic particles were clearly visible. Though only 75 measurable particles were found in one of the hand specimens (fig. 4-D') and 55 in another (fig. 4-D"), the fabric maxima were developed clearly. Both of them were northwest-southeast, though 50 degrees apart. It is difficult to remove pebbles from the matrix in Precambrian tillites, in order to check their elongated axes, unless the matrix is strongly weathered. Thus three dimensional orientation analyses, similar to those suggested by Dapples and Rominger for fine grained clastics (1945, p. 249), may give more reliable results than the above two dimensional measurements; but they are certainly more time consuming.

APPENDIX

The following procedure was used in the rapid till fabric studies of drill cores.

The 3¾-inch diameter cores were dried first, and then their orientation marks were deepened and colored. Then the cores were cut with a diamond saw in slices which were 1 to 1.5 inches thick. The cut surfaces were

brushed under a water spray for a minute and were then dried again. Brushing and washing removed about 1 mm of till matrix, thus exposing granules and pebbles as relief features (fig. 1).

For measuring compass directions of long axes, each slice of core was placed on paper in the middle of a circle (radius 3 inches), with their north-south orientation coinciding. An east-west line was also drawn through the centre of the circle. The upper half a circle was divided in ten-degree sectors. Measurement of the compass direction of each elongated pebble at least 2 mm long, as visible in the cut, was made with a transparent celluloid plaque, traversed by parallel lines (see figure 5). The plaque was placed on the till surface in such a manner that its lines were parallel to the apparent longest dimension of each pebble measured. That line which ran through the centre of the core gave a reading of the compass direction on the circle. This procedure was continued from one elongated particle to another, and each measured fragment was marked with a coloured pencil to avoid repetition. As the results were recorded by dots—one for each measurement—along the radii of the ten-degree sectors (see fig. 5), the maxima, if present, were seen at the completion of approximately 100 measurements, as in figures 2, 3, or 4.

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FIGS. 1–3, SEPTEMBER, 1959

NOTES

AGE OF INTERTIDAL TREE STUMPS AT WELLS BEACH AND KENNEBUNK BEACH, MAINE[1]

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Intertidal or subtidal tree stumps have been reported from several localities along the New England-Acadian coast. Lyons and Goldthwait (1934) describe stumps from Provincetown, Massachusetts; Scarboro, Maine; Grand Pre, Nova Scotia; Fort Lawrence, Nova Scotia; and Odiorne Point near Portsmouth, New Hampshire. Bradley (1953) describes intertidal stumps from Georgetown, Maine, and gives the age, as determined by radiocarbon analysis, as 4150 ±200 years. He also mentions an occurrence at Prouts Neck, Maine. To this list should be added the two occurrences at Wells Beach, Maine, and the two at Kennebunk Beach, Maine, which are described in this article. These two localities are in southern York County, approximately 32 and 26 miles south-southwest of Portland, respectively.

In the spring of 1955 following a severe northeast storm which caused extensive beach erosion, intertidal tree stumps were exposed on the crescent beach south of the center of town at Wells Beach. In a deep scour trough at mid-tide level on the beach, approximately 6 feet below high tide, 14 stumps ranging in diameter from 1 to 10 inches were found rooted in a layer of peat 1 to 2 feet thick. The peat is underlain by a very fine, gray, well-sorted sand of an undetermined thickness. One of these stumps was selected for radiocarbon analysis, the result of the analysis shown is in table 1. All radiocarbon ages reported in this article were determined at the U. S. Geological Survey Radiocarbon Laboratory, Washington, D. C.

In the same trough about 400 feet north of the stumps, a compact, iron-stained, and well-cemented till was exposed. The exact relationship of this material to the peat and underlying sand was not apparent because of the intervening veneer of modern beach sand; however, the till probably extends beneath the peat and sand.

As the result of another severe northeast storm during the spring of 1958, some of the stumps and peat were again exposed, but the till was not exposed.

In the Fall of 1954 Dr. Robert Dow of the Maine Department of Sea and Shore Fisheries located a large stump in a tidal channel in the marsh behind the northern beach at Wells Beach (Bradley, personal communication). A sample was taken by Bradley for radiocarbon analysis, and the result is shown in table 1. This stump was rooted in peat approximately 2½ feet below high tide. It evidently had been recently excavated from beneath 2½ feet of marsh sod by tidal currents and flood waters from a stream which empties into the channel. In addition, the writer has recently located several similar exposures in the same marsh. All are at approximately the same level.

During the summer of 1956 two occurrences of intertidal stumps were found at Kennebunk Beach, Maine. The western occurrence is in a small, rocky cove between Libbys Point and Lords Point (Wells 7½′ Quadrangle). The positions of 16 stumps were mapped and a sample was taken from one of them for radiocarbon dating (table 1). Some of the stumps are rooted in a thin layer of peat, but most seem to be rooted in silty clay. They occur from about three feet below mean low tide to mean low tide.

The eastern occurence is at the end of a beach just west of Oaks Point (Kennebunkport 7½′ Quadrangle). Here, several dozen stumps are exposed from about three feet above mean low tide to three feet below. They are rooted in a peat layer 2 to 2½ feet

[1] Manuscript received April 1, 1959.

TABLE 1.—*Radiocarbon dates of intertidal tree stumps at Kennebunk Beach and Wells Beach, Maine*

Location	U.S.G.S. Radiocarbon Laboratory no.	Wood type	Radiocarbon age (years)
Wells Beach	W 508	White Pine	2810 ± 200
Wells Marsh	W-396	White Pine	2980 ± 180
Kennebunk Beach West	W-509	Red Pine	1280 ± 200
Kennebunk Beach East	W-510	White Pine	3250 ± 200

thick. The result of a radiocarbon analysis of a sample from one of these stumps is given in table 1.

The ages of the stumps at Wells Beach and Wells Marsh are in very close agreement, supporting the inference drawn from field observation that these stumps grew contemporaneously and that the peat in which they are rooted is probably contiguous. The lower level of the beach occurrence may be due to compaction of the peat and underlying sand by the weight of the overlying beach deposits.

The date of the Kennebunk Beach East occurrence is generally of the same order of magnitude as those at Wells Beach and Wells Marsh, but the Kennebunk Beach West occurrence appears to be considerably younger, casting some doubt as to its validity. It is possible that this sample has undergone some contamination by recent C^{14}.

The ages of these stumps indicate that a lower sea level prevailed for an undetermined length of time prior to about 3000 years ago. During that time, trees grew over an area extending from the landward margin of the marsh to somewhere beyond the present extent of the beach. About 3000 years ago sea level rose, drowned these trees, and allowed beach deposits to migrate landward. Whether this change of sea level was eustatic or due to crustal movements is unknown; more radiocarbon dates for intertidal tree stumps from other parts of the coast are needed.

The writer wishes to thank Dr. Wilmot H. Bradley for permission to use the radiocarbon age of the stumps at Wells Marsh in this article and for radiocarbon analysis of other samples at the U. S. Geological Survey Radiocarbon Laboratory.

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CORRECTION OF SOME EARLIER DATA ON CALCITE AND DOLOMITE IN SEA WATER[1]

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Solubility products for calcite and dolomite in sea water reported earlier (Kramer, 1958) are in error and have been re-valued at: for calcite at 25°C, 0.65×10^{-6}, 1.40×10^{-6}, and 1.8×10^{-6} for 35, 45, and 60°/₀₀ salinity sea water respectively; and for dolomite at 25°C, 2.95×10^{-13}, 1.38×10^{-12}, and 2.04×10^{-12} for 35, 45, and 60°/₀₀ salinity sea water respectively. The data reported earlier for calcite at 5°C is questioned (due to possible temperature variation) although

[1] Manuscript received March 24, 1959.

no experimental checks have been made.

A free energy of formation value for dolomite (25°C) can be determined from the above data for calcite and dolomite and from the equilibrium constant for the dissociation of calcite at 25°C (Latimer, 1952, p. 319) of 4.7×10^{-9}. At constant temperature the difference in the equilibrium constant for calcite and the solubility product for calcite is due to the product of the activity coefficients concerned. Therefore,

$$\left(\frac{K_c}{K_c'}\right)t=\frac{(f_{Ca^{++}})(f_{CO_3^{=}})}{(f_{CaCO_3})} \qquad (1)$$

where K_c is the equilibrium constant for calcite at temperature t, K_c' is the solubility product (in sea water) at temperature t, and f is the activity coefficient of the constituent for molar concentration. If the activity of the solid phase is assumed to be unity, then

$$\left(\frac{K_c}{K_c'}\right)t=(f_{Ca^{++}})(f_{CO_3^{=}}) \qquad (2)$$

Also if the activity coefficients of the calcium and carbonate ions are of the same order of magnitude, the mean activity coefficient, $f\pm$, may be defined by

$$f\pm^2=(f_{Ca^{++}})(f_{CO_3^{=}}) \qquad (3)$$

The activity coefficient as regards carbonate chemistry is a function of the temperature and ionic strength (salinity) only; at 25°C, the mean ionic activity coefficients for sea water of 35, 45, and 60°/₀₀ salinity calculated from equations (2) and (3) are 0.085, 0.058, and 0.051 respectively. The free energy of dolomite is then calculated as follows:

for the reaction:

$$CaMg(CO_3)_2=Ca^{++}+Mg^{++}+2CO_3^{=}$$

1) $K_d=(a_{Ca^{++}})(a_{Mg^{++}})(a_{CO_3^{=}})^2$
$=(C_{Ca^{++}})(C_{Mg^{++}})(C_{CO_3^{=}})^2(f_{Ca^{++}})(f_{Mg^{++}})(f_{CO_3^{=}})^2$
$=$(solubility product)$(f\pm)^4$

2) $\Delta F^\circ=-RT\log K_d$
or $\Delta F^\circ=-1.364\log K_d$ (25°C)

3) $\Delta F^\circ=\Delta f^\circ_{Ca^{++}}+\Delta f^\circ_{Mg^{++}}+2\Delta f^\circ_{CO_3^{=}}-\Delta f^\circ_{dolomite}$
and $\Delta f^\circ_{dolomite}=\Delta f^\circ_{Ca^{++}}+\Delta f^\circ_{Mg^{++}}+2\Delta f^\circ_{CO_3^{=}}-\Delta F^\circ$

where K_d is the equilibrium constant for dolomite, a is the activity and C is the molar concentration for the constituents named, ΔF° is the free energy change (in kilo-calories) for the reaction, and Δf° is the free energy of formation (in kilo-calories) for the constituents named. Using the solubility products of dolomite and the activity coefficients of this paper, and the free energies of calcium, magnesium and carbonate ions found in Latimer (1952, p. 128, 316, 320), all at 25°C, the free energy of formation of dolomite is calculated to be 516.53, 516.54, and 516.61 K-calories for the three salinities. An average value of 516.5$_8$ corresponds to an equilibrium constant for solution of dolomite of 1.5×10^{-17} in agreement with that of Halla and Ritter (1935, p. 81) of 3×10^{-17}.

The new solubility products and free energy data completely negate the geological conclusions made earlier. According to the new data, dolomite is stable in sea water of normal composition with respect to calcite at 25°C. Figures 1a and 1b, derived from the solubility product data of calcite and dolomite and the calcium and magnesium ion concentrations in sea water, show the calcite and dolomite fields as a function of carbonate ion concentration and salinity, and pH and salinity. Dolomite should always develop over calcite since a smaller carbonate ion concentration is required for dolomite saturation compared with calcite. The probable reason that calcite forms instead of dolomite in recent marine sediments is due to the slow rate at which dolomite forms with respect to calcite. Calcite, however, is metastable and should alter to dolomite if the solution environment moved to the left of the calcite line (in fig. 1a or 1b) but within the dolomite field. Figure 1b shows that a small decrease in pH (with carbonate alkalinity/salinity constant) will cause calcite to dissolve and dolomite to develop. Dolomite appears to take a long time to form, and therefore some mechanism must ensure the constancy of carbonate ion concentration for dolomite development. An environment with a pH buffer could hold the pH (and carbonate ion concentration) constant for dolomite formation. The systems: NH_4OH—NH_4^+, $S^=$—HS^-—H_2S, CO_2—HCO_3^-—$CO_3^=$ or one of the weak organic acids could buffer sea water at a constant pH for the development of dolomite. Any weak acid or base with a dissocia-

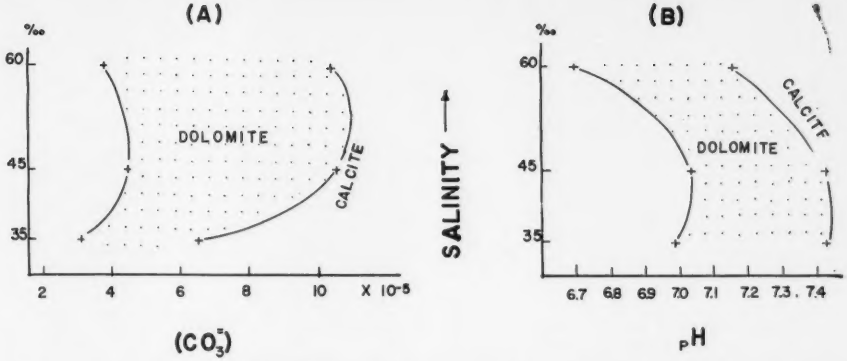

FIG. 1.—Equilibrium of calcite and dolomite in sea water (a) as a function of salinity and carbonate ion concentration; and (b) as a function of salinity and pH (assuming carbonate alkalinity/salinity $=$constant$=6.43\times10^{-5}$ equiv./liter °/₀₀). Dolomite stability and calcite instability is shown by the dotted fields.

tion constant near the value of the hydrogen ion concentration and with constituents in great excess compared to the hydrogen ion concentration would hold the pH constant.

Many naturally occurring buffering agents in different environments can be investigated with respect to calcite and dolomite formation in sea water. One possible mechanism of dolomite formation might be:

a) precipitation of calcium carbonate
b) restriction of the depositional environment during dry "warmer" periods increasing the activity of sulfur-reducing bacteria and decreasing and fixing the pH. ($S^{=}$, HS^{-}, $H_2S \gg H^{+}$)
c) development of dolomite or magnesian calcite from the calcareous ooze.

ACKNOWLEDGMENTS

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AUTHIGENIC TOURMALINE IN THE BANGANAPALLY STAGE (KURNOOL SYSTEM), INDIA

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During the detailed sedimentary studies of the rocks of Cuddapah and Kurnool Systems around Kalava, India (lat 15°37′ N.; long 78°12′ E.), authigenic overgrowths have been noticed on the tourmaline grains in pebbles of the conglomerate from the Banganapally Stage.

The Banganapally Stage is the lowest stage of the Kurnool System and the sequence starts with a basal conglomerate (50 feet thick) overlain by orthoquartzite. The conglomerate of this group is composed of pebbles of orthoquartzites, various coloured cherts, jasper, and hardened shales. The orthoquartzite pebbles in the conglomerates were seen to consist mainly of authigenic tourmaline which forms 93.4% of the bulk of the heavy minerals, with garnet (3.7%) and zircon (2.9%). About eight varieties of tourmaline have been identified and the deep brown variety (Fe-Schorlite) is the predominant type.

The authigenic growth is present on almost all the tourmaline grains. The overgrowths have taken place on the negative polar ends of the grains (1 and 2) and most of them are blue to colourless. The percentages of the different varieties of tourmaline (identified mainly on the basis of pleochroism and refractive index), with their optical characteristics are presented in table 1.

The deep brown variety has the maximum overgrowths, and the overgrowths are, in all the cases, pale blue to colourless with refractive index varying from 1.61 to 1.63.

Detrital tourmaline with authigenic overgrowths has been reported earlier by Alty (1933), Krynine (1946) and Gundu Rao (1952). The authors report, for the first time, such occurrences in the Banganapally Stage. According to Krynine, authigenic tourmaline generally forms at the bottom of the sea in a cold water environment, penecontemporaneously with the included sediment.

[1] Manuscript received April 1, 1959.

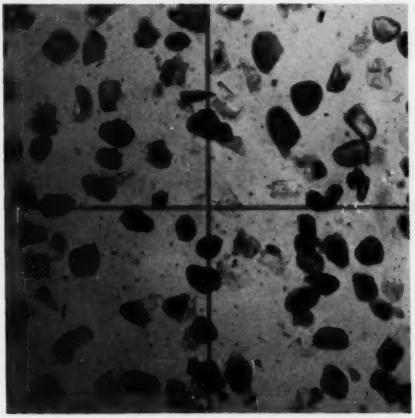

FIG. 1.—Photomicrograph showing different variaties of tourmaline grains with authigenic overgrowths (×35).

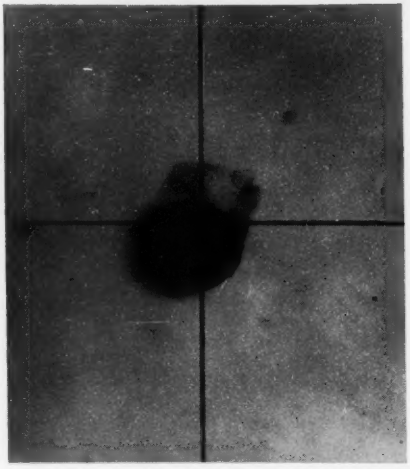

FIG. 2.—A well-rounded grain of authigenic tourmaline (at the centre) set in the position of maximum absorption showing overgrowth at one end of c-axis (×140).

TABLE 1.—*Varieties of tourmaline*

Pleochroic scheme		Refractive index		Variety	Percentage	% authigenic
z	x	ε	ω			
Black	Deep brown	1.65	1.70	Fe-Schorlite	36.0	57.3
Green	Pale brown	1.64	1.67	Schorlite	30.7	37.8
Deep blue	Pale pink	1.63	1.66	Indicolite	16.2	44.0
Pale blue	Colourless	1.62	1.64	Elbaite	5.3	18.9
Reddish brown	Pale brown	1.63	1.67	Fe-Dravite	5.1	31.00
Pink	Colourless	1.62	1.65	Li-rubellite	2.6	40.00
Deep Green	Yellowish green	1.64	1.68	Cr-tourmaline	2.1	—
Yellow	Colourless	1.61	1.64	Dravite	2.0	44.00

Because of its widespread occurrence even in completely welded orthoquartzites and non-porous calcareous sandstones, Krynine concluded that the process takes place at an early sedimentary age. Gundu Rao reported their occurrence in the Satyavedu Stage of Upper Gondwanas, near Madras.

The tourmalines reported in this paper occur in the orthoquartzite pebbles from the Banganapally conglomerate. The orthoquartzites generally form as blanket sands typically deposited under encroaching seas, after prolonged weathering, sorting and abrasion. The presence of authigenic tourmaline is much more widespread and can form under more varied environments than hitherto believed. The present case records its occurrence in a transgressive marine environment.

The ubiquitous presence of authigenic tourmaline in the Banganapally conglomerate zone is important as a good stratigraphic marker for correlation purposes.

The authors are thankful to Sri C. N. Rao for his suggestions.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 29, NO. 3, P. 470
SEPTEMBER, 1959

ANNOUNCEMENT

PALEONTOLOGICAL AND MINERALOGICAL ASPECTS OF POLAR WANDERING AND CONTINENTAL DRIFT—A SYMPOSIUM

Atlantic City, April 25–26, 1960

Over 30 years ago, the American Association of Petroleum Geologists held its memorable symposium on continental drift. Coincidentally, this anniversary was marked by the publication of two symposia in widely separated parts of the world—"Continental Drift, A Symposium" held at the University of Tasmania in March 1956, was published at Hobart, Tasmania, 1958, and "Polar Wandering and Continental Drift, A Symposium" was published in the Journal of the Alberta Society of Petroleum Geologists, Vol. 6, Nos. 6 and 7 (June and July) 1958.

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